Climatic Feedbacks and Desertification: The Mediterranean Model

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ABSTRACT

Mesometeorological information obtained in several research projects in southern Europe has been used to analyze perceived changes in the western Mediterranean summer storm regime. A procedure was developed to disaggregate daily precipitation data into three main components: frontal precipitation, summer storms, and Mediterranean cyclogenesis. Working hypotheses were derived on the likely processes involved. The results indicate that the precipitation regime in this Mediterranean region is very sensitive to variations in surface airmass temperature and moisture. Land-use perturbations that accumulated over historical time and greatly accelerated in the last 30 yr may have induced changes from an open, monsoon-type regime with frequent summer storms over the mountains inland to one dominated by closed vertical recirculations where feedback mechanisms favor the loss of storms over the coastal mountains and additional heating of the sea surface temperature during summer. This, in turn, favors Mediterranean cyclogenesis and torrential rains in autumn–winter. Because these intense rains and floods can occur anywhere in the basin, perturbations to the hydrological cycle in any part of the basin can propagate to the whole basin and adjacent regions. Furthermore, present levels of air pollutants can produce greenhouse heating, amplifying the perturbations and pushing the system over critical threshold levels. The questions raised are relevant for the new European Union (EU) water policies in southern Europe and for other regions dominated by monsoon-type weather systems.

1. Background and introduction

Around the Mediterranean Sea, deserts and desert-like conditions can be found in close proximity to a very warm sea and thus to a marine air mass with a high moisture content, for example, the coasts of eastern Tunisia, Libya, Egypt, and the province of Almeria in Spain. These regions were covered with vegetation in historical times, for example, during the Roman Empire (Bölle 2003a). In Almeria, the dense oak forests covering its mountains were cut to fuel lead mines just 150 yr ago (Charco 2002), and the area became the desert-like site for the spaghetti western films. The question is, Did these areas become desertified (or was the desertification accelerated) by removing the forests and desiccating the coastal marshes?

According to Claussen (2001), an issue of climate feedbacks is that "predictions of global atmospheric models are highly sensitive to prescribed large-scale changes in vegetation cover" and that "although available studies illustrate the potential effects of massive vegetation changes on the climate system, they can hardly be validated." We feel, however, that the characteristics of the atmospheric circulations in the western Mediterranean basin, resulting from its latitude,

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orography, and land–sea distribution, make this region an ideal test ground for checking whether or not vegetation is a passive component of climate and probably for other questions related to feedbacks in climate modeling.

In Spain, the increase in mountain agriculture began in the late 1500s, and extended grazing began in the early 1700s. These transformations included cutting down forests, clearing large pasture areas, and terracing mountain slopes, all of which increased erosion. The growth of the Ebro River delta began, in fact, in the sixteenth and seventeenth centuries (Puigdefábregas and Mendizabal 1998). The accumulated changes had already led to marginal (i.e., subsistence) agriculture by the time of the Spanish Civil War (1936–39), which was followed by a massive population exodus from the rural areas to the larger Spanish cities (Madrid, Barcelona, Valencia, and Bilbao) lasting from the mid-1940s to the late 1960s. Lands were abandoned, limited grazing remained, and, by the time the trend slowed down in the late 1970s, only the very elderly were left in many villages that have become, de facto, "ghost towns." These migrations have contributed to desertification in Spain (Puigdefábregas and Mendizabal 2003) and, through similar mechanisms, in other Mediterranean countries (Bächler 1995).

For the past 2000 yr, and especially during the nineteenth and twentieth centuries, coastal marshes around the Mediterranean Sea have been drained and desiccated, initially for health reasons, that is, to avoid malaria, and later for agricultural use (Bolle 2003a; b). As the coastal areas have also been at the receiving end of the migrations mentioned (three of the cities above are coastal), they have become increasingly urbanized and industrialized since the late 1950s. The rate of change increased greatly after the 1974 Middle East war and was followed by extended demands for housing (residential, vocational, and tourist). As a result, much of the coastal agricultural land used in the 1950s has been urbanized, and few (or none) of the virgin coastlands remain. Finally, along with the increased anthropogenic pressures, other disturbances in this region, for example, air pollution and forest fires, have also increased significantly in the last three decades (Millán et al. 1998; Moreno and Oechel 1994; Moreno 1998).

The study of atmospheric processes in the Mediterranean basin has been part of the European Commission's Research Programs, and authors of this paper have participated (since 1974) in projects related to atmospheric chemistry (air pollution meteorology), desertification and carbon balances (see appendix). In fact, it was while collecting field information for planning measurement campaigns (in 1986–87) that members of the team were first told that storms in the mountains surrounding the Mediterranean Sea did not reach the mature-storm stage as frequently as in the past. And, according to local senior residents, the creeks where they had often fished in their childhood were now dry most of the year. Similar information has been conveyed to us since then, regarding the loss of summer storms in the Italian provinces of Basilicata and Calabria, as well as in Sardinia, Sicily, and other Mediterranean islands.

The purpose of this paper is to combine experimental and modeling results from the aforementioned projects with precipitation data to derive working hypotheses about the possible processes and feedbacks involved in the changes observed in the summer storm regime. Section 2 describes the precipitation types and presents the criteria used to disaggregate the main components from the daily rainfall data series in the study region. Section 3 reviews the most relevant findings regarding the sea breezes and other mesoscale winds that lead to the vertical recirculations and long residence times of air masses within the western Mediterranean basin. Section 4 uses experimental data and modeling results to estimate the temperature and the moisture gained by the sea breeze on its way inland, as well as the additional temperature increases resulting from the radiative effects of air pollution. The individual and combined effects of these factors on the cloud condensation level (CCL) are also evaluated in section 4, while section 5 addresses the possible feedbacks. Finally, section 6 presents the conclusions.

2. Precipitation types and data analysis

a. *Precipitation types*

The study area considered (Fig. 1) includes the Valencia region and the neighboring inland areas reached by the combined sea breeze and upslope winds, that is, the combined breeze (section 3). Extended use of daily recording precipitation gauges in rural Spain began in the 1950s, and the number of stations in the study area has increased from 60 in 1950 to 509 at present. Of these, 60 have climatologically homogeneous records (i.e., continuous data for $\geq$30 yr) since 1950, and 250 stations now cover the last 30 yr (to 2001). For this work, 116 stations were selected with a 97% data confidence level. However, even though the dataset is climatologically homogeneous, the spatial distribution of the stations reflects the site criteria of more than 30 yr ago.

The average precipitation series (yearly precipitation total collected by selected stations divided by number of selected stations each year) is shown in Fig. 1. The linear regression shows a decreasing trend and, in spite of the increase in the number of stations over the years, the average series becomes more spiky toward the end. The 5-yr running average for the same data series (Fig. 2a) smoothes out the individual peaks but shows a deepening of the large cycles with the maxima increasing and the minima decreasing during the period considered.

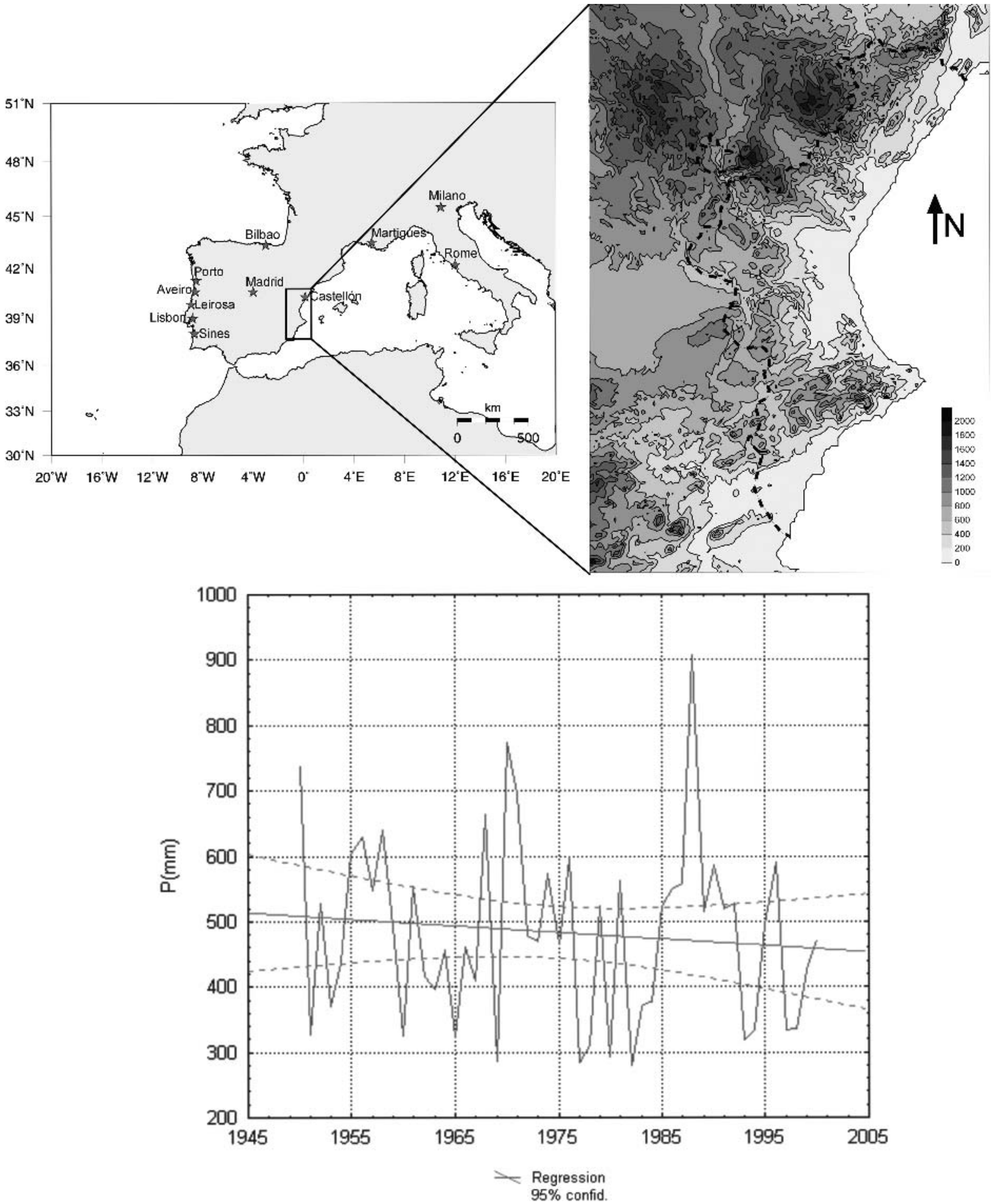

FIG. 1. Location of the study area within Europe, the Mediterranean basin, and area-averaged precipitations for the Valencia region for the period 1950–2001. Stars mark sites where meteorological measurements, including simultaneous vertical sounding, were performed during the measurement campaigns.

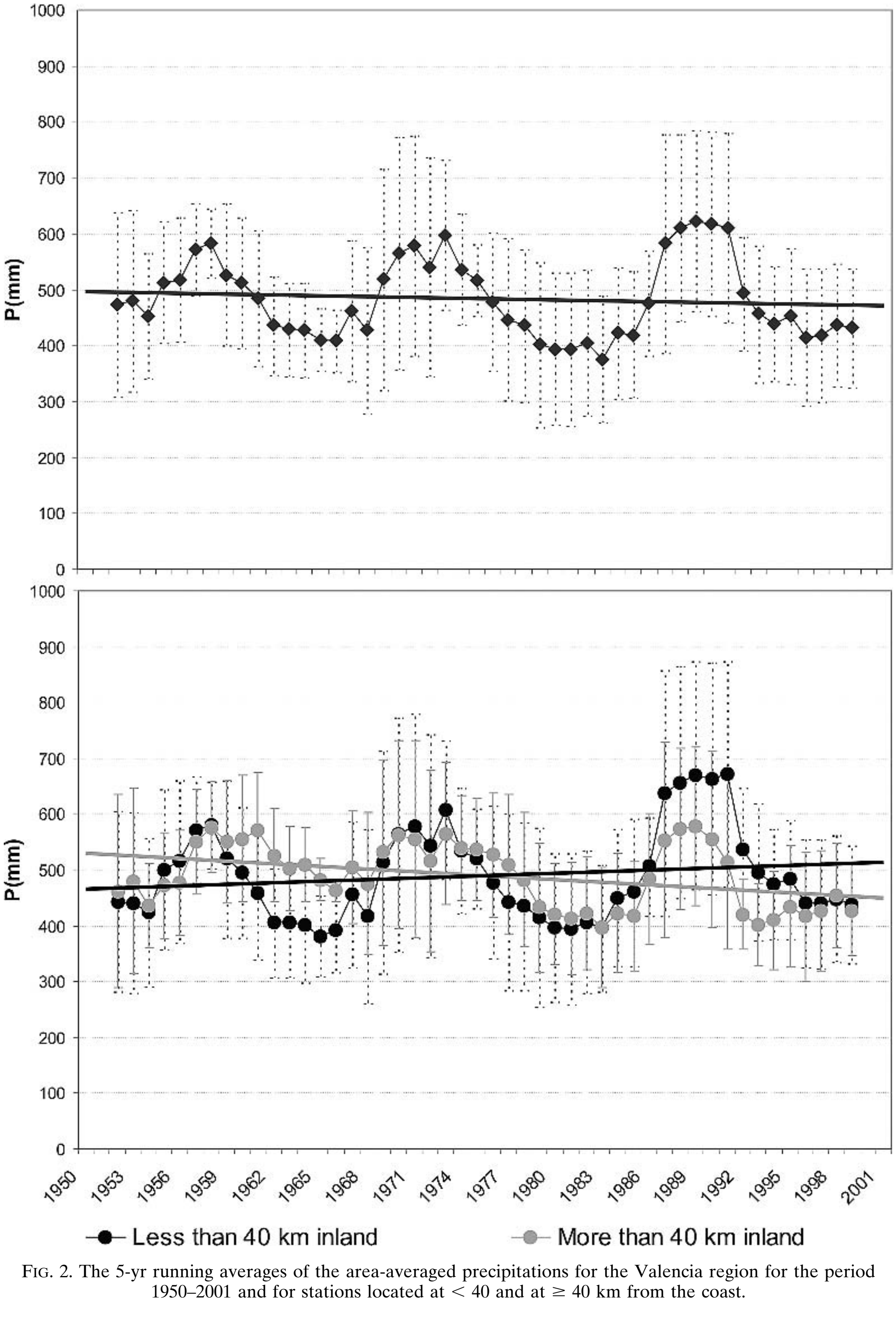

FIG. 2. The 5-yr running averages of the area-averaged precipitations for the Valencia region for the period 1950–2001 and for stations located at < 40 and at ≥ 40 km from the coast.

This picture changes when the data are separated according to the distance from the station to the coast (Fig. 2b). For the stations in the coastal plains located at <40 km from the coast, the linear regression shows an increase in precipitation and the 5-yr running average accentuates the increase in the maxima (see below). For the stations in the mountains inland at ≥40 km from the coast, the regression shows a marked decreasing trend, and it is the minima that become smaller in the 5-yr running average.

A meteorological analysis of the precipitation events shows that the rain regimes in this region are clearly distinct spatially and temporally (Estrela et al. 2000; Peñarrocha et al. 2002). The last orographic barrier encountered by Atlantic fronts before moving into the Mediterranean is the coastal mountain chain. Precipitation from these fronts occurs mainly on the windward (west) side of the mountains (Fig. 3a) and is followed by a Föhn-type effect over the Spanish east coast (Millán et al. 1998). Summer storms, on the other hand, are associated with the final stages in the development of the combined breeze, and the precipitation occurs mainly in the mid- to late afternoon on the east side of the same mountain ranges (Fig. 3b).

Intense precipitation in the western Mediterranean occurs mainly in autumn–winter when cold(er) mP or cP air masses are advected over a warmer sea, and the moisture recharge is governed by the differences in temperature between the sea surface and the air mass. This mechanism is highly sensitive to changes in the Mediterranean sea surface temperature (SST; Millán et al. 1995; Pastor et al. 2001), but the areas affected by the precipitation depend on the direction of the advected air mass (Meteorological Office 1962; Barry and Chorley 1987).

Advections from northwest to north along the Carcassone Gap and/or the Rhöne valley reinforce cyclogenesis to the lee of the Alps (Genoa depressions), which affects mainly Italy and parts of northern Africa (Tunisia). The western coasts of Mediterranean North Africa, the Spanish east coast, and southern France are affected by northeasterly to southeasterly advection, which we consider as examples of backdoor (cold) fronts (Huschke 1959; Bluestein 1993). Figure 3c shows the resulting spatially averaged pattern.

b. Precipitation data analysis

Table 1 includes the main characteristics and the meteorological criteria used by our group to disaggregate the daily precipitation data into the three main components. However, because the generalized availability of synoptic weather maps for Spain dates from 1959, the detailed treatment had to be limited to the period 1959–2001. For each day in this period, the criteria were applied to the joint analysis of the daily precipitation data; the surface pressure maps at 0000, 0600, 1200, and 1800 UTC; and the 500- and 300-hPa maps at 1200 UTC. Figure 4 shows the results for the subset of stations on the coastal strip (<40 km from the coast) and for the stations in the inland mountains (≥40 km from the coast). And thus, bearing in mind the caveat about the length of the records, the disaggregated data indicate that in the last 41 yr, the following conditions are true:

- The average annual precipitation over the coastal strip has not changed significantly.
- On the other hand, the average annual precipitation over the stations inland shows a decreasing trend.
- At present, Atlantic fronts contribute approximately 20% of total precipitations. Their contribution shows a decreasing trend in both subareas, though more pronounced inland.
- Summer storms contribute approximately 11% of the total precipitation, and their contribution shows a decreasing trend over the whole area (coastal strip and inland).
- Backdoor fronts contribute approximately 65% of total. The average for this component has remained essentially unchanged over the mountain areas but shows an increasing trend over the coastal strip. The major torrential events in this region, with large floods in 1959, 1972, 1980, and 1989 appear in this series.
- Finally, since summer storms and intense precipitation originate mainly from water evaporated from the Mediterranean Sea, their sum, that is, ≈75%–80%, could be considered an estimate of the precipitation generated by (self-) evaporation within the region.

The perceived change in the storms inland, therefore, seems to be the result of a decreased contribution from Atlantic (cold) fronts and summer storms, while on the coastal strip the precipitation losses from summer storms and Atlantic fronts appear to be balanced by an increase in the contribution from backdoor (cold) fronts. This situation has been further analyzed, and the results in Fig. 5 confirm a trend suggested by Figs. 1, 2, and 4b, that is, that precipitation in the region has also changed in character, becoming more torrential, that is, more precipitation in a smaller number of larger events occurring closer to the coast. Figure 5 also shows an increase in large events during the first part of the year.

The decreases in the contributions by frontal systems probably derive from changes in the conditions over the Atlantic and are not analyzed here. Finally, calculations using climatological data and experimental results from the European Commission (EC) projects (Table 2) show that the air mass in the combined breeze transports enough water vapor to feed the equivalent of several summer storms each day. The questions then are as follows: why do summer storms appear to be diminishing, and why are torrential rains increasing?

3. Mesoscale atmospheric circulations: The experimental and modeling evidence

Atmospheric processes in the Mediterranean region vary with the season and specific basin (Meteorological Office 1962; Barry and Chorley 1987). The eastern Mediterranean (sub)basin is limited to the south by a mountainless desert. In summer, it falls under the influence of the Asian monsoon system, and advection dominates (Kallos et al. 1998). At the same time, the western Mediterranean (sub)basin, totally surrounded

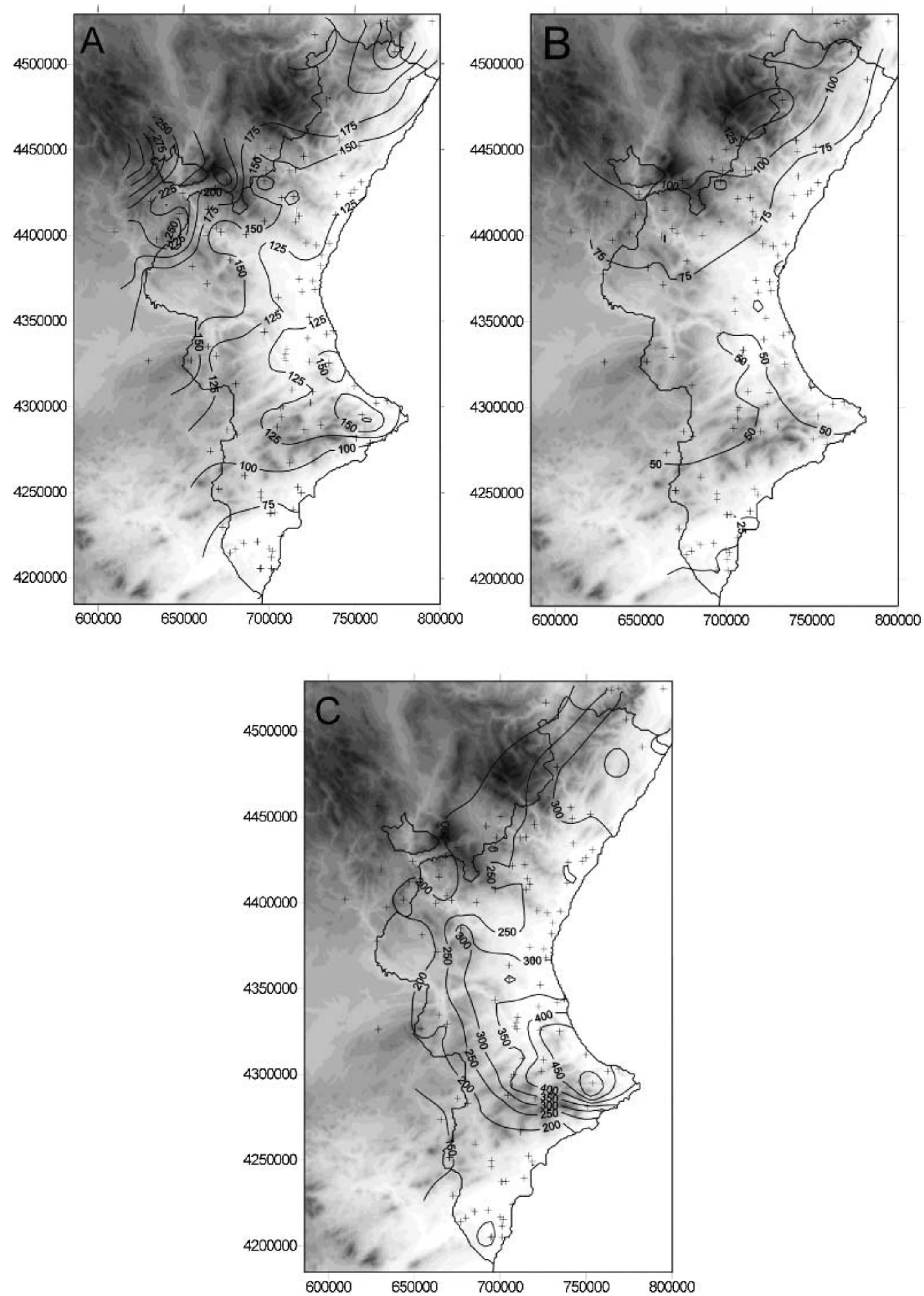

FIG. 3. Averaged spatial distribution of precipitation types for the period 1959–2001. Contributions from (a) Atlantic fronts, (b) summer storms, and (c) backdoor (cold) fronts (Mediterranean cyclogenesis). Crosses mark the stations used.

TABLE 1. Precipitation types in the Valencian region.

Component	Characteristics	Criteria
Precipitation from classic (Atlantic) fronts	Their most frequent occurrence is from early autumn to late spring.	Collected rain amounts were assigned to this category whenever a clearly noticeable frontal passage had occurred during the precipitation event.
Summer storms	In this category we consider the storms driven by the combined sea breeze and upslope winds.	Precipitation was attributed to this category whenever the Iberian Thermal Low was observed at 1200 and/or at 1800 UTC on the day of the event.
	They tend to develop during the afternoon over the top of the coastal mountain ranges, some 60–80 km inland, and are more frequent from late Apr to Sep.	
	During the late evening and night, the storms migrate easterly (toward the coast) before they dissipate.	
Backdoor cold fronts	Precipitations associated with Mediterranean cyclogenesis. They can be very intense and occur mainly over the sea and coastal areas and near mountain slopes in autumn and winter.	To be included in this category, the precipitation event must coincide with easterly advection over the western Mediterranean, that is, a well-established anticyclone over central Europe, accompanied but not necessarily limited by the development of a low-pressure area over northern Africa and/or a pool of cold air aloft over the Iberian Peninsula.

by high mountains, falls under the influence of the semipermanent Azores anticyclone. Clear skies prevail under a generalized level of subsidence aloft, and mesometeorological processes with marked diurnal cycles dominate (Millán et al. 1997).

a. Processes at the meso-γ and meso-β scales

In the western basin, the EC research projects documented that the upslope winds reinforce the sea breezes along the coasts and create "combined breezes," known to be stronger than the sum of their components (Mahrer and Pielke 1977; Miao et al. 2003). In these circulations the sea breeze progresses inland during the morning and early afternoon in a stepwise fashion by incorporating weaker upslope circulatory cells (Millán et al. 1992; Millán et al. 1998; Salvador et al. 1997). By late afternoon, the sea-breeze front tends to become locked along the ridges of the mountains surrounding the basin (Millán et al. 2000; Millán et al. 2002).

Deep vertical injections take place at the front of the combined breeze, and the mountain slopes can be considered to act as convective–orographic chimneys that link the surface flows directly with their return flows aloft. By the time the circulations are fully developed, several return layers can be observed (Fig. 6), and vertical injections higher than 3–4 km over the ridges of the coastal mountains 60 to 100 km inland have been documented (Millán et al. 1992) and modeled (Salvador et al. 1997; Salvador et al. 1999; Millán et al. 2002). The return flows produce a system of stacked layers over the sea (Fig. 7), which leads to vertical recirculations of pollutants emitted along the coasts. Experiments on the Spanish east coast during the MECAPIP project (see appendix for definitions of EC projects hereafter) showed that a tracer emitted into the sea breeze (i.e., inland bound) reentered inland coming from the sea 2 days later (Millán et al. 1992).

Figure 7 also shows the temperature profile measured along leg 1 of the flight (highlighted). The layer from ≈2000 to 3000 m follows, essentially, a dry adiabat with a potential temperature of ≈315 K, and the profile can be compared with two Regional Atmospheric Modeling System (RAMS) simulations ($\boldsymbol{\Delta}$). In Fig. 7b, the monthly averaged western Mediterranean basin SST distribution was used, and in Fig. 7c, the SST was obtained from National Oceanic and Atmospheric Administration (NOAA) satellite measurements on the day previous to the measurements. The simulated profiles compare well with observations but do not reproduce the finer temperature structure associated with the (air pollution) layers.

b. Processes at the meso-α scale

Another finding during the RECAPMA project was that the combined breezes, their return flows aloft, and their compensatory subsidence over the sea become self-organized by late morning at a scale that encompasses the whole western basin. The main features of this flow are illustrated in Fig. 8, that is, the western basin becomes like a large cauldron that boils from the edges toward the center. The final result is that a regionally linked vertical recirculation system is generated (Millán et al. 1997). Similar effects have been documented in other areas of the Mediterranean (Kallos 1997; Kallos et al. 1998).

When the coastal recirculations (1–2-day return times; see above) grow and become part of the larger one (regional scale) the results are increased stratification and longer recirculatory times. The motion of multiple particles has been simulated using the Hybrid Particle and Concentration Transport Model (HYPACT;

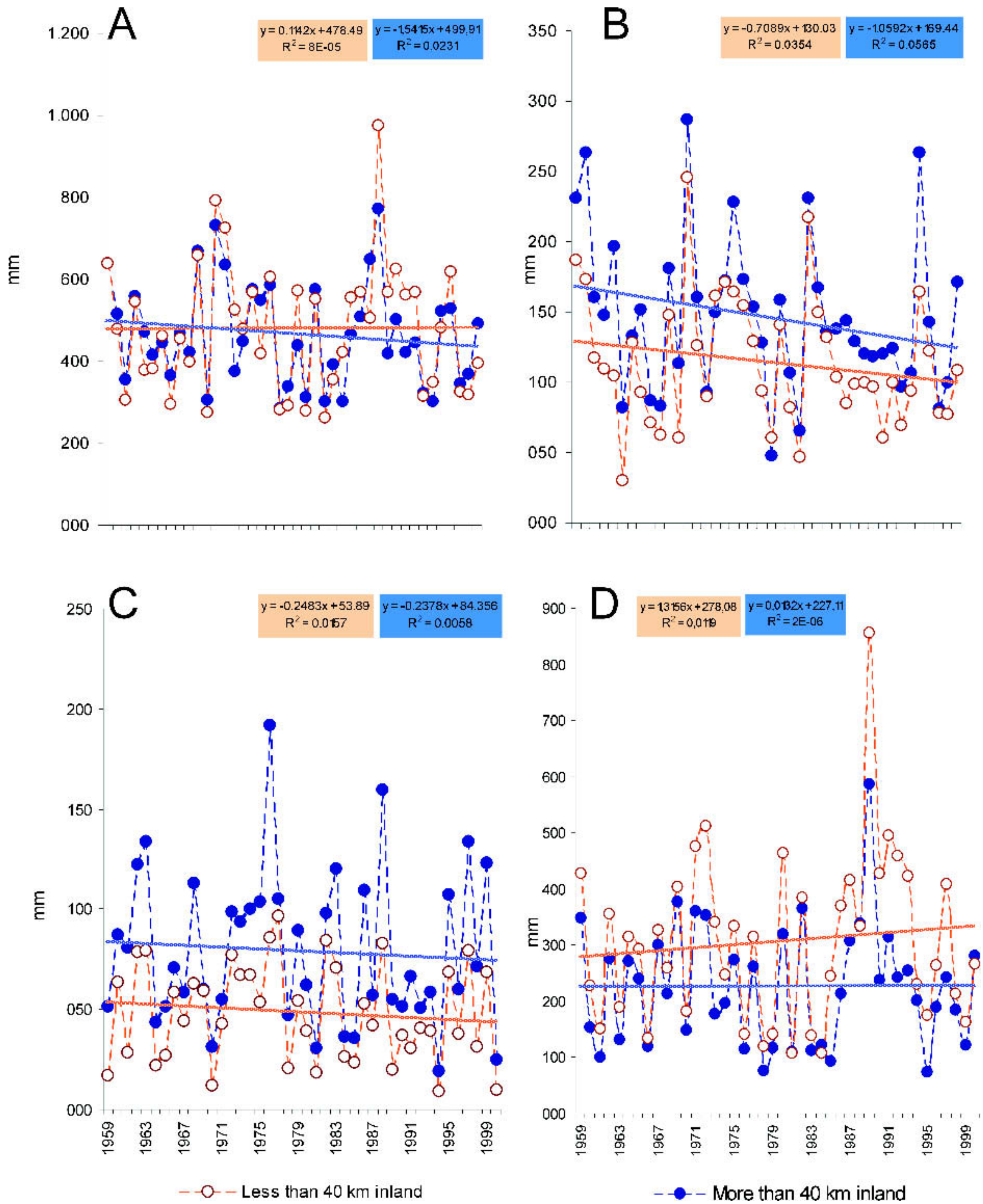

FIG. 4. Time series of the area-averaged precipitation for the subsets of stations located at < 40 and ≥ 40 km from the coast, for the period 1959–2001: (a) precipitation total, (b) Atlantic fronts, (c) summer storms, and (d) backdoor cold fronts.

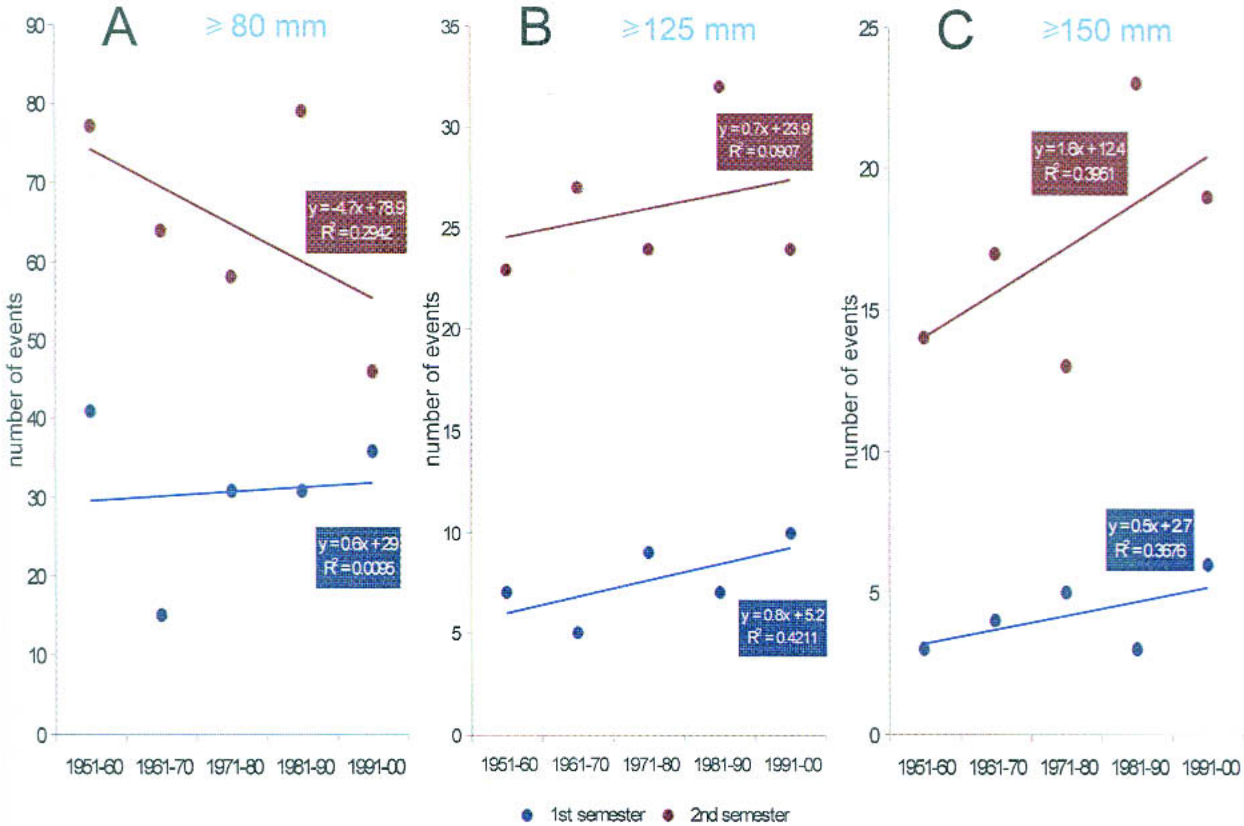

FIG. 5. Decadal variation in the number of intense rain events during the first and second parts of the year for the period 1951–2000, for amounts (a) ≥ 80, (b) ≥ 125, (c) ≥ 150 mm day^{-1}.

Tremback et al. 1993) coupled to the meteorological fields provided by either RAMS or the fifth-generation Pennsylvania State University (PSU)–National Center for Atmospheric Research (NCAR) Mesoscale Model (MM5; Gangoiti et al. 2001). The results indicate that the time period for renovating 80% of the air mass below approximately 3500 m in the western Mediterranean basin in summer is of the order of 7 to 10 days, depending on the point of release. In this process, a particle can enter inland with the sea breeze, be lifted into the return flows along the orographic chimneys in the breeze front, sink over the sea, reenter inland with

TABLE 2. Precipitation estimates based on experimental and climatological data.

Applicable results from European research projects (western Mediterranean basin) ■ Average dewpoint (D_p) temperatures at coastline: 18.2°C (Jul), 20.2°C (Aug). ■ Average sea surface temperature: 26°C. ■ Average distance from coast to top of coastal mountain ranges: ≈80 km. ■ Average distance traveled by combined breeze: ≈160 km day^{-1}. ■ Final depth of surface boundary layer: ≈250 m. ■ Observed range of dewpoint temperatures at 78 km inland: 22°–28°C. ↘	Typical storm ● 5×10^8 liters rainwater, that is, 20 L m^{-2} over a footprint of 2.5 km wide by 10 km long. ↓ Air volume displaced by combined breeze ● Slab of air 10 km wide along the coastline (= to length of storm track considered), times ● Average depth for the combined breeze ≈250 m, gives → 2.5×10^6 m^2, times ● Average traveled distance 160 km, yields ● Displaced air volume: 4×10^{11} m^3 (≈4.78×10^{11} kg) ↘	Available water vapor at beginning of breeze run ● Assume an average value of 19°C Dp, that is, a mixing ratio ≈13.6 g kg^{-1} (≈11.4 cm^3 m^{-3}) of water vapor at the coast ↓ Further assumptions ● Only half of the displaced air mass (i.e., 80 km long) participates in the precipitation, ● Only 5 g kg^{-1} of the water vapor initially available at the coast is effectively rained out, ↓ Potential precipitation ● 10^{12} cm^3 of water (≈ twice the 5×10^{11}cm^3 required for a typical storm).

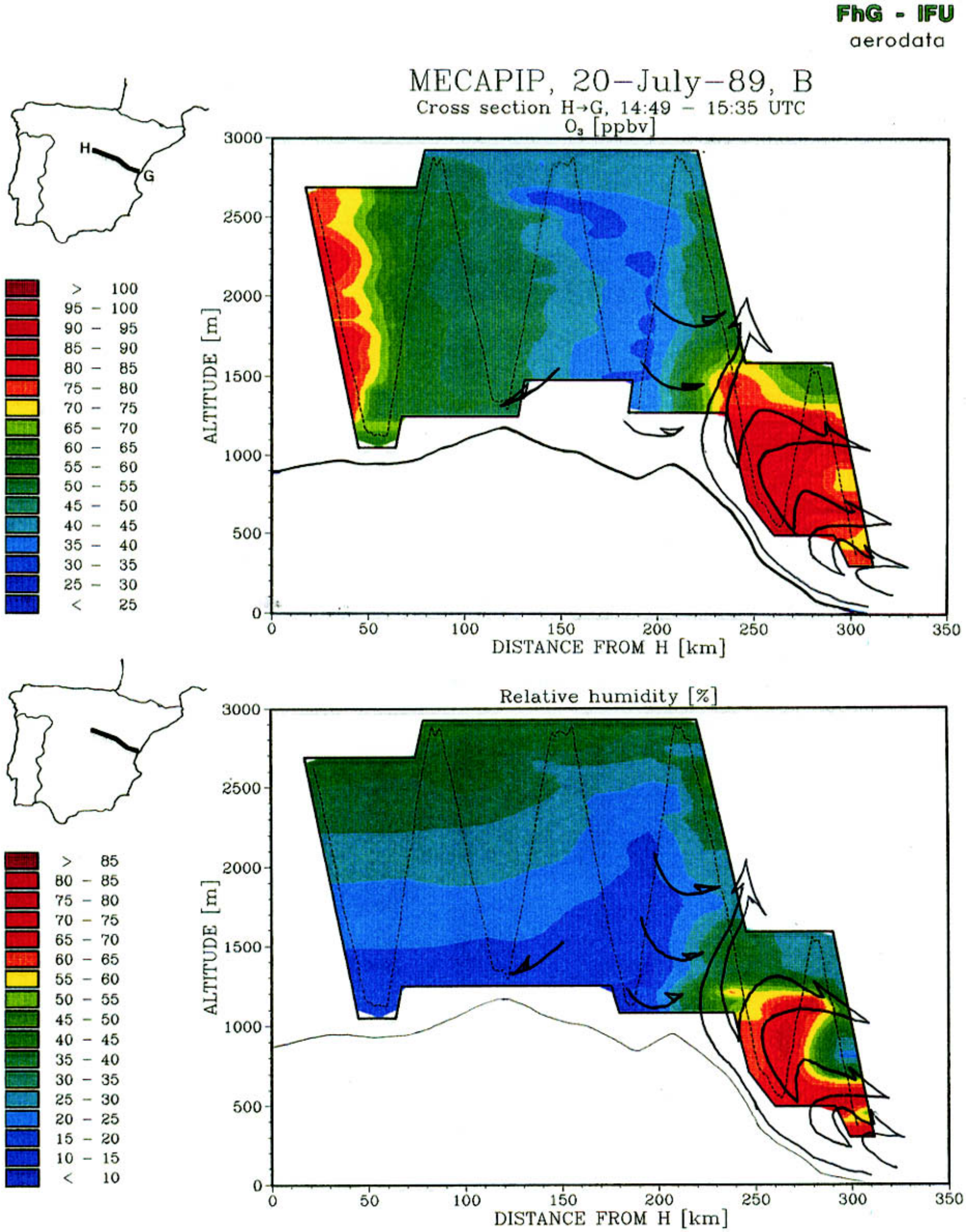

FIG. 6. Development of vertical recirculations on the Spanish east coast, as documented by the ozone and relative humidity profiles obtained by an instrumented aircraft at 1449–1535 UTC on 20 Jul 1989. The actual flight path (seesaw pattern) is shown with a dotted line, and the ground track relative to the Iberian Peninsula appears at left.

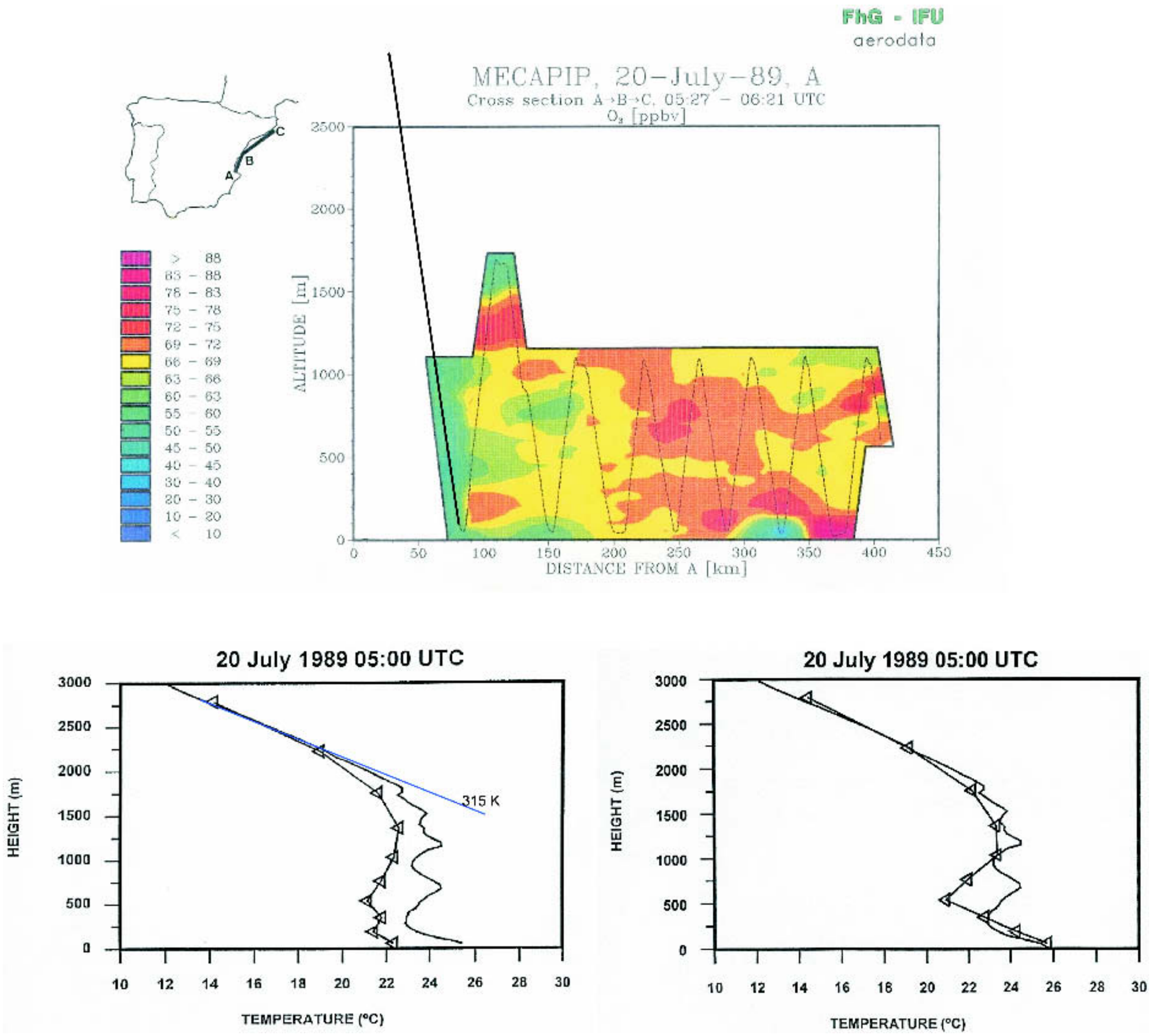

FIG. 7. (a) Vertical ozone distribution along 450 km of the Spanish east coast at 0527–0621 UTC on 20 Jul 1989, showing the layered structure resulting from the recirculations on the previous day(s). The aircraft flew in a seesaw pattern over the sea (track A–B–C at left) at 40–60 km from the coast. (b), (c) Measured, and simulated (Δ), temperature profiles along leg 1 of the flight (highlighted); see text.

later, and return aloft toward the sea and do this several times before escaping out of the basin. Just for comparison purposes, during summer the air masses below ≈3500 m over the United Kingdom, under mainly advective conditions, can be renovated once or twice a day.

c. The effects of compensatory subsidence on sea-breeze development

Finally, as the local-to-regional circulations grow and become self-organized at the regional level during the day, their compensatory subsidence intensifies over the sea (Figs. 8 and 9) and over the coastal areas. This generates a lower inversion along the coasts, which has been observed to sink during the afternoon and confine the inland-bound leg of the surface flows to a depth of ≈150 to 300 m (Millán et al. 1992; Millán et al. 1997; Millán et al. 2002). Thus, by the time the regional flows are fully developed, this lower inversion caps the combined breeze and extends most of the way from the coast to the mountains surrounding the basin. A very significant finding, therefore, is that per unit distance along the coasts, the combined sea breezes and upslope winds end up displacing a volume of air that is very long (≈160 km) but rather shallow (on the average ≈250 m deep in eastern Spain).

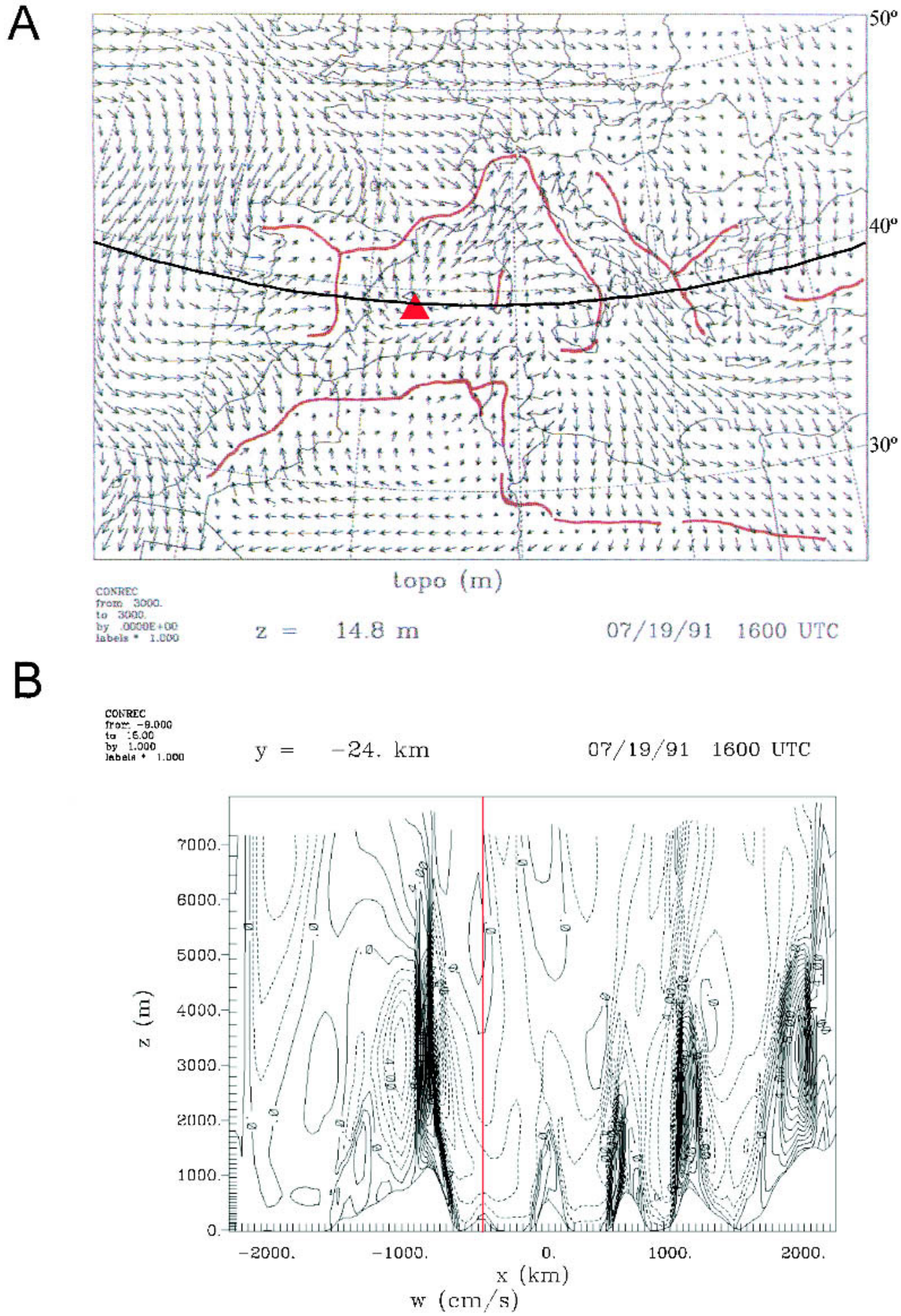

FIG. 8. RAMS simulation for the Mediterranean at 1600 UTC on 19 Jul 1991. The western basin shows an anticyclonic-like field of surface winds (14.8-m height) flowing toward convergence lines on the coastal mountain ranges (red lines). (b) Vertical speeds over a parallel (39.5°N) passing near the base of the spiral flight (red triangle) for Fig. 9; ascent is represented by solid line and sinking is represented by dotted line. The simulated subsidence near the Balearic Islands (at ≈ -4 cm s^{-1}) greatly underestimates the measured values (see Fig. 9).

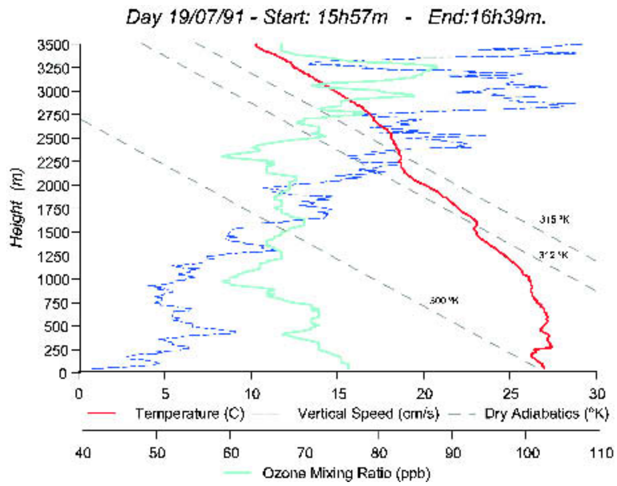

FIG. 9. Experimental profiles of ozone, temperature, and vertical speed obtained over the Balearic basin (marked point in Fig. 8) from 1557 to 1639 on 19 Jul 1991. The data show an average O_3 concentration in the order of 70 ppbv ($\approx$140 μg m^{-3}) through the sounding. The measured sinking speeds near the top ($\approx$3500 m asl) are in the order of 15 to 25 cm s^{-1}. The structure of the temperature profile, with long and essentially adiabatic sections, is indicative of sustained sinking of air masses that have acquired equal potential temperature.

The grid sizes in the mesometeorological models used (RAMS and MM5) have been optimized to incorporate the effects of the complex terrain surrounding the Mediterranean basin (Salvador et al. 1999). Additional methods have also been developed to incorporate the real Mediterranean sea surface temperature distribution derived from NOAA satellite measurements (Bádenas et al. 1997). With these, we have been able to simulate the main features of the observed wind fields, including the deep vertical exchanges along the mountains, and also the precipitation fields produced by torrential rains (Pastor et al. 2001). However, even using the finest vertical resolution at the limit of numerical instability, we still have not been able to reproduce the amount of compensatory subsidence observed (cf. Figs. 7, and 8, and 9), nor the fine structure of the resulting vertical stratification (Millán et al. 2000; Millán et al. 2002) nor the inversion capping the sea breezes. Thus, we feel that the real residence recirculation times could be even longer than those mentioned above.

d. Implications for climatological studies

The key characteristics of the summer atmospheric circulations in the western Mediterranean basin are 1) the narrow and deep vertical exchanges along the mountain ranges surrounding the basin, which produce stratified layers over the sea; 2) the self-organization of the coastal circulations at the regional scale; and 3) the resulting compensatory subsidence of these layers over the sea, during the day, as part of a basin-scale vertical recirculation. And, for the purpose of this work, the most important aspects are 4) the vertically confined nature of the surface flows in the coastal recirculations, 5) the long residence times of the air masses involved, and 6) the resulting accumulation of pollutants (ozone), as shown in Figs. 7 and 9.

A relevant aspect relating to climate is that air masses involved in the vertical recirculations can acquire a memory of their previous interactions with the surface (land and sea), for example, heating, moisture-content gains. This suggests that feedback processes within the surface–atmosphere–hydrological system could manifest themselves much earlier in the western Mediterranean than in regions where air masses are in transit most of the time.

Given the limitations of current mesoscale models, a helpful situation for analyzing feedbacks derives from the vertically confined nature of the surface flows and the almost laterally homogeneous layered structure over the sea resulting from the vertical recirculations (Fig. 7). The air mass in the combined breeze can thus be assumed to maintain its integrity along its path over the surface; that is, lateral exchanges are compensated by air with similar characteristics, while the surface air mass remains essentially free from vertical exchanges until it reaches its orographic chimney inland. This means that the simplest approach to evaluate possible feedbacks involving exchanges of the surface flows is an atmospheric thermodynamic diagram.

4. Surface exchanges and air pollution effects

a. Surface heating

The temperature gained by the surface air mass after traveling over the sun-heated coastal plains and slopes can be obtained from the temperature data in the upper part of the observed return flows. To exclude additional gains due to latent heat exchanges from condensation, the profile in Fig. 7 was used because no cloud formation had been observed over the mountains on the previous days. The potential temperature of the upper layer, that is, $\approx$2000 to 3000 m, is 315 K and indicates that by the time the air mass that leaves the coast at $\approx$26°C (299 K) becomes injected into the return flows, it has gained $\approx$16 K.

Gains of this order are consistent with most of the experimental temperature profiles obtained during the measurement campaigns in this region in summer, and the 312–315-K potential temperature range appears to be clearly associated with the upper part of the return flows in the coastal recirculations (see also Fig. 9). If there are no other processes involved, Fig. 10 indicates that the cloud condensation level (CCL) resulting from the interception of the 315-K potential temperature line and the mixing ratio line corresponding to a dewpoint of 19°C at the coast (the average value for July and August) would occur above $\approx$2740 m, which is substan-

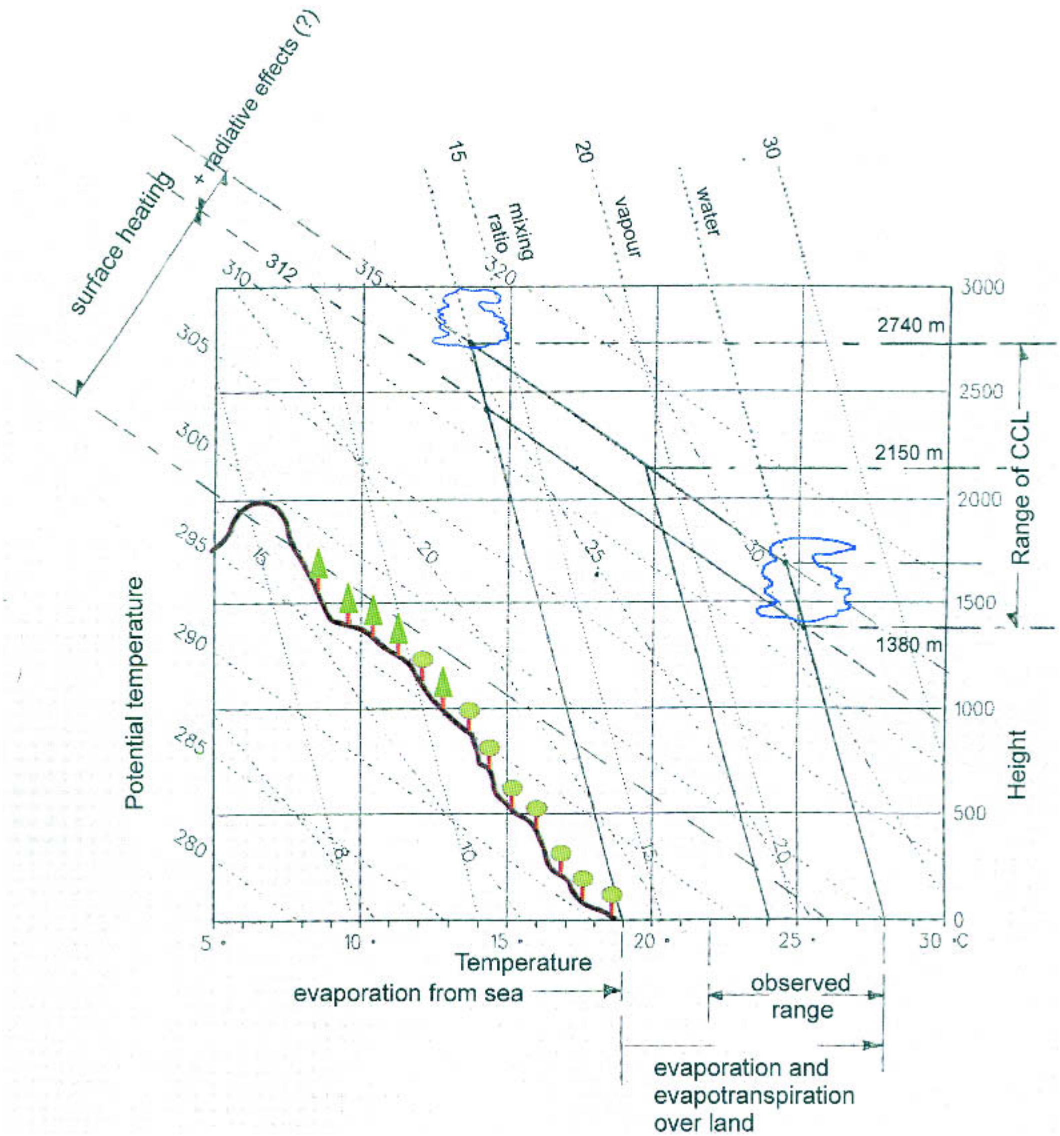

Fig. 10. Emagram showing possible cloud condensation levels for the conditions and processes observed in the region.

tially higher than the coastal mountains in this region (1800 to 2060 m).

b. Evapotranspiration and land-use changes

To calculate the water vapor contributed by vegetation to the combined breeze we assume 1) that the combined breeze displaces a volume of air 160 km long by 250 m high, per unit length along the coast; 2) that evapotranspiration takes place only along the first 80 km of its path, as a typical distance from the coast to the mountain tops; and 3) that evapotranspiration takes place during the breeze period. Finally, estimates of the actual evapotranspiration for this region in summer vary from 5 to 7 (+) $\mathrm{Lm^{-2}\ day^{-1}}$ over the irrigated coastal plains, to 1–3 $\mathrm{Lm^{-2}\ day^{-1}}$ over the drier areas (pines and maquia) in the mountains inland, and we have further assumed a value of 6 $\mathrm{lm^{-2}\ day^{-1}}$ along the first 20 km and 2 $\mathrm{Lm^{-2}\ day^{-1}}$ along the remaining 60 km.

We find that vegetation contributes $\approx$5 g $\mathrm{kg^{-1}}$ of additional water vapor to the volume of air displaced by the combined breeze. Thus, by the time the air parcel reaches the mountain ridges, its water vapor mixing ratio has increased to $\approx$19 g $\mathrm{kg^{-1}}$ (a dewpoint of $\approx$24°C), from the initial 13.8 g $\mathrm{kg^{-1}}$ (19°C dewpoint) at the coast. This value compares favorably with the 22°–28°C range of dewpoint temperatures measured during the MECAPIP (Millán et al. 1992) and follow-up projects at the Valbona-Mora site located 78 km inland.

Figure 10 indicates that the CCL for the estimated value of the mixing ratio and the 315-K potential temperature drops by some 600 m, bringing it down to approximately 2150 m, that is, closer to the mountain peaks. The CCL drops to 1700 m if the highest observed value of the dewpoint temperatures (28°C) is considered, and even further, to ≈1400 m, if the 312-K potential temperature is assumed as a lower estimate for heating the surface air mass.

The range of possible heights for the CCL, that is, from 1400 to 2750 m, indicates that both heating and evaporation along the surface play major roles in the formation of summer storms in these circulatory systems; the range also shows how sensitive the system is to variations in these two components. The moisture added by evaporation over the coastal wetlands and by the vegetation along the breeze path lowers the CCL and can be considered to act as a priming/trigger mechanism that favors the onset of condensation/precipitation, tending to "flush away" part of the water vapor already available in the marine air mass. Thus, increasing this moisture contribution would propitiate the development of a feedback mechanism toward more storms and more vegetation.

On the other hand, any perturbation in the surface characteristics that further heats the surface air mass, for example, increasing the surface dryness, would lift the CCL and lower the probability of storms developing. The same result would occur by diminishing the sources of evaporation and evapotranspiration, for example, drainage of coastal marshes and/or changes from flooding to drip irrigation procedures, increased deforestation, and/or forest fires, or combinations of these.

c. *Air pollution effects*

Under strong summer insolation, the coastal recirculations become large natural photochemical reactors where coastal emissions of NO_x and other precursors are transformed into oxidants, acidic compounds, aerosols, and ozone. Observations on rural mountain sites in the region show average daily O_3 values within the 125–160 $\mu g\ m^{-3}$ range almost every day from May to August (Millán et al. 2000; Millán et al. 2002), and aircraft measurements over the Mediterranean Sea (Figs. 7 and 9) also yield average values of 70 ppb (140 $\mu g\ m^{-3}$) within the lower 3500 m.

Satellite data (Fishman et al. 1990) yield vertically integrated O_3 values over the western Mediterranean of the order of 50–55 Dobson units during June–August, that is, an average of 125–150 $\mu g\ m^{-3}$ within an 8000-m-high column. Finally, modeling of the ozone surface concentrations (Bastrup-Birk et al. 1997; Lelieveld and Dentener 2000) yields comparable averages (120–140 $\mu g\ m^{-3}$), thus placing observations and modeling results within the same overall values.

The observed tropospheric ozone, however, is always associated with haze, and a number of research projects have been proposed to the EC (e.g., PHAMA) to determine their origin, composition, vertical structure, radiative effects, and fate. Some of the relevant questions regarding this matter are whether aerosols reflect solar radiation and cause cooling (Lelieveld et al. 2002) or whether, as in the case of contrails (Travis et al. 2002), the combined radiative effect of aerosols mixed with pollutants (and/or water vapor) showing greenhouse properties is to produce additional heating.

Actual indications of the net radiative effects in the western basin can be obtained by comparing the temperature profiles derived from high-resolution mesoscale modeling with the measured temperature profiles associated with the layers of ozone and aerosols, as shown in Fig. 7. The two simulations run with RAMS were intended to explore a range of possibilities since the net radiative effect could, in fact, be sea surface heating (see below).

In both cases, the model matches the upper part of the observed temperature profile but shows lower values in the lowest 2200 m. In the first case, the differences increase toward the bottom and reach an average value of approximately 2–3 K throughout the lower 1500 m. In the second, the lowest boundary values match, and the differences between simulation and measurements are most marked at the peaks associated with the pollution layers. These differences increase from ≈0.7 K at 1750 m to more than 3 K between 500 and 750 m.

Given the observed temperature range in the return flows, the 312-K line can then be considered the result of surface heating alone, and the 315-K line can be considered the present limit of accumulated surface heating plus (possible) radiative effects from air pollution. Thus, if the net radiative effect of present tropospheric ozone and aerosol levels is to heat the lower atmosphere by an additional 1–3 K, the CCL of the air mass in the combined breeze would rise some 100 to 300 m by the time it reaches the mountain ridges. And raising the CCL higher over the mountain tops would lower the probability of the storms maturing.

5. Feedbacks and runaway effects

When storms mature, some of the water vapor in the combined breeze is rained out, part of the released latent heat is used to feed deep convection, and another part is advected away at the tropopause level. In these conditions, the surface air mass becomes entrained throughout the troposphere, and the coastal circulations can be considered "open." This would be the situation in a monsoon-type circulation. In the western Mediterranean, the transition from a vertical recirculation in the morning to an open-type circulation with the development of moist convection, cumulus, and storms in the afternoon still occurs frequently on summer days along the Italian Appenine Mountains (Cantú and Gandino 1977; Millán et al. 1997).

If the water vapor is not condensed, Fig. 6 shows that it follows the ozone into the return flows of the combined breeze. Thus, while this situation lasts, the coastal wind system remains "closed," that is, dominated by orographically aided convection that favors the vertical recirculations. Raising the CCL of the surface air mass in this region, therefore, provides a mechanism for keeping the coastal circulations closed and for loading the troposphere with additional moisture to the depth reached by the coastal recirculations, that is, to approximately 2500–3500 m over the sea.

Finally, the reevaporation of droplets from the unmatured cumulus clouds also increases the water vapor content in the layers above $\approx$3500 m. How this contribution compares to the moisture reevaporated during the dissipating phase of a matured storm and the resulting temperature profiles through the troposphere for either case are two additional questions. It is significant that through the mechanisms described, moisture loading and stability changes can take place through the troposphere under anticyclonic conditions of subsidence aloft.

It is at this point that the self-organizing nature of the coastal recirculations and the resulting long residence times within the western basin come into play. The additional moisture could contribute to the creation of an anomalously deep, moist, and potentially unstable troposphere awaiting the right mechanism to trigger precipitation, for example, destabilization by advection of cold air aloft. Whether or not this can explain the increase in large storms over the coasts and sea in spring is a question that must still be determined. In any case, the more humid air mass may also be advected away and participate in stronger precipitation elsewhere [e.g., 2002 summer floods over Central Europe Ulbrich et al. 2003a,b)].

Moreover, water vapor is an important (the most important?) greenhouse gas, and increasing its amount in the midtroposphere over this region would add to any radiative heating effects caused by air pollutants, leading to additional rises in land and sea surface temperatures and, ultimately, in the temperature of the air near the surface. An increase in evaporation from the sea would tend to lower the CCL of the airmass in the combined breeze. Increased heating by the land surface (with lower heat capacity than the water) would tend to raise the CCL. Which process dominates, and in which order, are two more questions that must be answered.

Additional heating of the sea surface is cumulative and can lead to delayed effects. For example, it may result in higher SST in the western Mediterranean basin by early autumn and winter and favor more intense rains (Millán et al. 1995; Pastor et al. 2001). Radiative heat trapping may also result in delayed cooling of the sea surface in winter and faster warming in early spring and may explain the shift in the observed frequency of torrential rains shown in Fig. 5.

6. Discussion

Available experimental results suggest that, because of the nature of the atmospheric circulations along its coasts, the hydrological cycle in the western Mediterranean basin has been drifting toward a critical threshold level, that is, when the cloud condensation levels of the surface air masses rise above the coastal mountain ridges, leading to the loss of summer storms. Any additional perturbations to the surface air masses, including increasing their temperature (by whatever means) or decreasing evaporation and evapotranspiration along their paths, or any combinations of these, could tip the present situation further toward the loss of mountain storms and desertification at the local–regional scale.

The question of the end radiative effect of air pollution in the Mediterranean basin, that is, is it cooling or warming the sea and surface air mass, is still open, even though our results indicate that warming dominates in the western Mediterranean at this stage. The hypothesis that the radiative effects of particulates cool the Mediterranean (Lelieveld et al. 2002) would indeed lead to less evaporation and fewer summer storms, but it would also lead to fewer torrential rains in autumn–winter–spring, which is contrary to observations in the western basin.

Our working hypotheses on the possible reasons for the observed precipitation changes in the western Mediterranean basin relate to 1) land-use changes, as the main factor drying the surface and 2) atmospheric composition changes, that is, an increase in tropospheric ozone and aerosols producing additional greenhouse heating. The hypotheses are as follows:

- Land use that results in surface drying leads to increased heating of the surface air mass, lower evaporation, higher cloud condensation levels, and thus fewer summer storms in the mountains inland.
- The radiative effects of air pollution could be further heating the surface air mass and lifting the CCL; this could be perturbing (present) local balances just enough to tip the equilibrium toward decreased summer storms.
- In turn, the recirculatory nature and the long residence times of the air masses in this region lead to the accumulation of ozone and water vapor (both acting as greenhouse gases) in the lower-to-upper troposphere over the western Mediterranean, adding to the heating of the sea surface during the summer and to an increase in torrential rains in autumn and winter at the regional (basin) scale.

In relation to climate–vegetation feedbacks we can consider the following:

1) long-term effects, that is, fewer storms in the mountains, positive feedback through vegetation losses (less evapotranspiration), and increased surface heating (drier soil) toward desertification;

2) delayed effects, that is, a warmer Mediterranean by the end of summer and early autumn leading to more frequent and intense torrential rains, by positive feedback. These rains can occur *anywhere in the basin* and result in flash floods over the coasts and nearby mountain slopes; and
3) if the slope vegetation response to these flash floods is increased soil erosion, then the positive feedback toward desertification would be further reinforced.

Finally, another far-reaching conclusion is that any perturbation to the hydrological cycle in any part of the western basin can propagate to other parts of the basin and neighboring regions. This derives from the fact that the warmer water pools move (i.e., rotate) within the basin, and, thus, any increase in sea surface temperature can result in torrential rain and flash floods anywhere in the basin.

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APPENDIX

Definitions of EC Project Acronyms

MECAPIP	Meso-Meteorological Cycles of Air Pollution in the Iberian Peninsula (1988–91)
RECAPMA	Regional Cycles of Air Pollution in the Western Mediterranean Area (1990–92)
SECAP	South European Cycles of Air Pollution (1992–95)
T-TRAPEM	Transport and Transformation of Air Pollutants on East Mediterranean (1992–95)
BEMA	Biogenic Emissions in the Mediterranean Area, Phase I (1993–95) and Phase II (1996–98)
MEDEFLU	Carbon and Water Fluxes of Mediterranean Forest and Impacts of Land Use/Cover Changes (1998–2000)
RECAB	Regional Assessment and Modelling of the Carbon Balance within Europe (2000–03)
ADIOS	Atmospheric Deposition and Impact of Pollutants, Key Elements and Nutrients on the Open Mediterranean Sea (2000–03)
CARBOMONT	Effects of Land Use Changes on Sources, Sinks and Fluxes of Carbon in European Mountain Areas (2001–04)
PHAMA	Photo-Oxidant and Aerosols in the Mediterranean Area (1999)

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