Climatology and Dynamics of the Summer Etesian Winds over the Eastern Mediterranean*

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ABSTRACT

The Etesians are persistent northerly winds that prevail over the eastern Mediterranean during summer. A climatology of Etesian outbreaks over the Aegean was compiled with the aid of the 40-yr ECMWF Re-Analysis (ERA-40) dataset and their vertical organization is investigated. Their variability arises from high-frequency variability originating in the midlatitudes, interannual and intraseasonal variability controlled by the South Asian monsoon, and a local diurnal cycle. Consistent with the monsoon influence, Etesian outbreaks are most frequent from mid-July to mid-August. In agreement with previous studies, a negative trend in the incidence of Etesian outbreaks is detected during the overall June–September period, which is strikingly strong for September but diminishes in June. The strengthening of the Etesians by day over the central and southern Aegean results from the deepening of the Anatolian thermal low because of the daytime sensible heating near the surface. The timing of an outbreak onset is controlled by wave disturbances originating over the North Atlantic that trigger the development of a strong ridge over the Balkans, which induces anomalously strong northerly flow and subsidence over the Aegean. During Etesian outbreaks, sharp tropopause folds and stratospheric intrusions of high potential vorticity descend deeply into the troposphere.

1. Introduction

The summer circulation over the eastern Mediterranean (EM) is dominated by persistent northerlies (Fig. 1), known since antiquity as the Etesians (HMSO 1962; Metaxas 1977; Maheras 1980; Reddaway and Bigg 1996; Lolis et al. 2002). This name derives from the Greek word “etos” (year) to emphasize their annual recurrent nature. They arise as a sharp east–west pressure gradient prevails over the EM during summer owing to a pressure dipole between the Persian trough (Bitan and Saaroni 1992; Saaroni and Ziv 2000; Alpert et al. 2004; Saaroni et al. 2010), which emanates from the massive Asian heat low

to form the Anatolian thermal low, and high pressure over central Europe and the western Mediterranean. The low-level cold-air advection induced by the Etesians moderates the heat and human discomfort in the region (Metaxas and Bartzokas 1994; Koletsis et al. 2009). However, their ventilating effect is largely balanced by adiabatic warming induced by large-scale subsidence (Tyrlis et al. 2013, their Fig. 3) that inhibits cloud formation and results in clear skies (Ziv et al. 2004), a typical feature of the Mediterranean climate in summer.

The Etesians control many aspects of the EM climate. The southward sea wave activity in the Aegean and the Levantine basin is associated with the persistent wind stress exerted on water volumes by the Etesians (Lionello and Sanna 2005), which can also force distinct sea currents and eddies over the entire Levantine basin (Zecchetto and De Biasio 2007; Amitai et al. 2010) that in turn affect marine and coastal environments. Episodes of strong Etesians can spawn forest fires that degrade air quality and damage the Mediterranean terrestrial ecosystems (e.g., Amiridis et al. 2012). The Etesians regulate air quality over the EM during the summer when boundary layer pollution standards are often exceeded owing to the transport of emissions from the European industrial zones

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FIG. 1. July climatology of MSLP (shaded) and 1000-hPa horizontal wind (arrows). The profiles presented in Figs. 3 and 4 refer to the flow averaged over the small boxes (solid lines) presented here and in Fig. 2. The seasonal cycles illustrated in Fig. 14 refer to PV averaged over the Aegean and the large boxes (dashed lines). The coordinates of all areas are listed in Table 1.

and cities (Kallos et al. 1998; Lelieveld et al. 2002; Kalabokas et al. 2008) and from biomass burning around the Black Sea (Sciare et al. 2003; de Meij and Lelieveld 2011). On the other hand, the Etesians block the northward transport of desert dust toward the EM, while local pollution in urban environments is dispersed to levels typical for rural areas (Theodosi et al. 2011).

The recurrent nature of the EM circulation is attributed to the climatological stability of the monsoon signal. This influence consists of a large-scale background state with a baroclinic signature comprising an upper-level ridge and a low-level trough (Persian trough) that begins to expand westward in late spring and early summer. Dynamically, it consists of an upper-level warm structure and subsidence areas, which are associated with equatorially trapped and westward-propagating Rossby waves excited by the South Asian monsoon convection (Rodwell and Hoskins 1996, 2001). Steep sloping isentropes, with a northwest–southeast orientation (Tyrlis et al. 2013, their Fig. 10), over the EM and the Middle East facilitate further subsidence on the western and northern periphery of the warm structure, which is exposed to the midlatitude westerlies. The monsoonal influence on the EM circulation is demonstrated by the strikingly synchronous July maxima of the Etesians' intensity and subsidence over the EM with that of monsoon convection over northern India, where the weak easterly upper-level jet favors a strong Rossby

wave response and impact on the midlatitude circulation. The dynamics of Etesians are tightly interwoven with the large-scale subsidence observed over the EM (Tyrlis et al. 2013). The large-scale background circulation induced by the monsoon each summer controls the stable appearance of the Persian trough and thus regulates the contribution of the eastern low-pressure center to the establishment of the pressure gradient over the EM that favors the Etesians. Many previous studies highlight the importance of the anticyclonic center over the Balkans (Metaxas and Bartzokas 1994; Kassomenos 2003; Burlando 2009; Poupkou et al. 2011) and its role in establishing the strong pressure gradient and the Etesians over the EM.

The pronounced topography over the EM and the Middle East amplifies the large-scale monsoon-induced circulation. Orographically locked summer anticyclones enhance the mid-/low-level northwesterly flow on their eastern flanks (Tyrlis et al. 2013, their Fig. 3b), leading to distinct subsidence maxima, such as over the EM. The northerly flow is sustained and accelerated through topographic channels, such as over the Aegean Sea, where the wind speed often exceeds 15 m s^{-1} , with gusts over $20\text{--}25 \text{ m s}^{-1}$ (Kotroni et al. 2001; Burlando 2009). Climatologically, the Etesians are northeasterly over the northern Aegean and northerly over the central and southern Aegean, while they become northwesterly farther southeast (Fig. 2). The northeasterly and northwesterly flow, respectively, over the aforementioned areas

FIG. 2. Mean JA wind at (a) 1000, (b) 925, (c) 850, and (d) 700 hPa. Wind speed (m s^{-1} ; shaded) is also plotted. The grid points included in the calculation of mean wind speed and direction over the Aegean are marked with dots only in (a). The calculations are also performed separately for the northern (filled rhombuses) and southern (filled circles) Aegean. The coordinates of AEG, NAE, and SAEG are listed in Table 1.

contribute to the characteristic overall ring shape of the Etesians around the Anatolian Peninsula that extends from the Black Sea to Cyprus, delineating the extension of the Middle East thermal low over Turkey. The Etesians are significantly weaker over the surrounding land areas, while they retain some strength over the Ionian Sea, especially over its southern parts, though they blow from a northwesterly direction. Over the Adriatic Sea, the weak northerly flow interacts with the sea–land breezes (Klaic et al. 2009), while farther south the Etesians blow uninterrupted toward the Sahel monsoon system (Raich et al. 2003). In the vertical direction, the characteristic topographically forced ring shape fades above 850 hPa (Fig. 2) where the effect of topography on the flow diminishes. Although the flow acquires a more westerly direction with increasing height, it retains a northerly component up to around 500 hPa (Tyrilis et al. 2013, their Fig. 12a). Thus, the Etesians extend aloft and they are not merely a near-surface phenomenon. The

interaction of the flow with the complex topography in and around the Aegean can lead to modification of the flow with funneling effects between islands that can be hazardous for ship navigation or significant flow deceleration at the upwind side of larger islands, such as Crete (Brody and Nestor 1980; Kotroni et al. 2001; Koletsis et al. 2009, 2010).

The Etesians over the Aegean are characterized by recurrent periods of gale-force northerlies interrupted by quieter spells. In the sparse related literature, the term Etesians refers either to these episodes or to the background circulation, which can be confusing (Repapis et al. 1978). Here, we identify the spells of enhanced Etesians, referred to as Etesian outbreaks. Semiobjective methods have been utilized for the selection of a day of Etesians following the inspection of daily mean sea level pressure (MSLP) charts to identify cases with strong pressure gradients over the Aegean (Prezerakos 1984). Statistical methods have also been employed to study the

importance of the anticyclonic center over central Europe for the regime of Etesians (Metaxas and Bartzokas 1994). Wind observations from the National Observatory of Athens were used to define a day of Etesians when the northerly flow over Athens supersedes the local southerly sea breeze (Carapiperis 1951, 1960; Repapis et al. 1978). The total number of days of Etesians in each month [parameter L in Repapis et al. (1978)] was found to be a reliable metric of the frequency of Etesians over the Aegean. This parameter was later exploited by Poupkou et al. (2011) for the “tuning” of a wind-speed threshold in order to recognize a day of Etesians based on National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) gridded data.

This study expands the analysis by Tyrlis et al. (2013) concerning the influence of the South Asian monsoon over the EM on seasonal time scales. It pursues the understanding of the midlatitude influence on the western anticyclonic center over central Europe that in turn controls the onset of Etesian outbreaks. The development of an index quantifying such an important climatic phenomenon is beneficial for the understanding of the present and future climate of a region, which is projected to be a climate change “hot spot” (Lelieveld et al. 2012). In particular, the study of the influence of midlatitude dynamics on short time scales requires the construction of a daily index describing the Etesians. Section 2 presents all the steps followed for the definition of such an index and describes the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) data used in this study. Section 3 presents aspects of a climatology of the Etesians and studies their variability on various time scales. Specifically, on subdaily time scales, the diurnal cycle of the phenomenon and its driving dynamics are investigated. Furthermore, the seasonal cycle of the phenomenon and its interannual variability are studied. Section 4 analyses the dynamical precursors that are responsible for the higher-frequency variability of the phenomenon leading to the onset of Etesian outbreaks. Section 5 investigates the mechanisms explaining the persistence of the Etesians. A novelty of this study is that the analysis is not confined near the surface, but it explores the full vertical structure of the Etesians throughout the troposphere. This description culminates in the detailed study of the upper-level disturbances and it provides evidence for the development of tropopause folds around the onset of Etesian outbreaks, which are discussed in section 6. Also, this section expands the results by discussing the dynamical links among the Etesians, the tropopause folds, and the contribution of the stratospheric intrusions of ozone-rich air to the observed summer ozone pool over the EM. Finally, section 7 summarizes the results and presents conclusions.

2. Data and methodology

a. Data description

The analysis is based on the ERA-40 dataset (Uppala et al. 2005). Analyses of 6-hourly ERA-40 MSLP, geopotential height, temperature, and wind components are used, which are available on an N80 full-Gaussian grid ($\sim 1.1^\circ \times 1.1^\circ$) and on pressure levels spanning the depth of the troposphere. Effectively, 45 summers during the period 1958–2002 are included in our analysis. Following Tyrlis et al. (2013), the thermodynamic energy equation terms are quantified during the ERA-40 period, which facilitates the study of the Etesian outbreaks dynamics (section 5) and their diurnal cycle (section 3). The equation can be written in isobaric coordinates (Holton 1992) as follows:

$$\underbrace{\frac{\partial T}{\partial t}}_{\text{LTT}} = \underbrace{\frac{J}{C_p}}_{\text{TPTT}} - \underbrace{\frac{T}{\theta} \frac{\partial \theta}{\partial p} \omega}_{\text{VATT}} + \underbrace{\left(-u \frac{\partial T}{\partial x}\right)}_{\text{ZATT}} + \underbrace{\left(-v \frac{\partial T}{\partial y}\right)}_{\text{MATT}}, \quad (1)$$

$\underbrace{\hspace{10em}}_{\text{HATT}}$

where T is temperature, θ is potential temperature, u and v are the horizontal wind components, ω is the vertical velocity (omega), J is the diabatic heating rate, and C_p is the specific heat of dry air at constant pressure. The term $\partial T/\partial t$ corresponds to the local temperature tendency (LTT), J/C_p represents the total physics temperature tendency (TPTT), $-(T/\theta)(\partial\theta/\partial p)\omega$ gives the temperature tendency due to vertical advection (VATT), whereas the terms $-u(\partial T/\partial x)$ and $-v(\partial T/\partial y)$ represent the temperature tendency due to zonal advection (ZATT) and meridional advection (MATT), respectively. The sum of the terms ZATT and MATT corresponds to the total temperature changes caused by horizontal advection (HATT). Thus, TPTT is directly calculated from the 6-hourly ECMWF model forecasts, as in Tyrlis et al. (2013).

b. Index describing the intensity of Etesians

In this study, daily indices are devised to quantify the mean intensity of the northerlies over the Aegean. The methodology involves the temporal averaging of the wind speed and direction (on the chosen isobaric level) during a calendar day, at every grid point, and the subsequent spatial averaging for the grid points shown in Fig. 2a that delineate the area affected by the Etesians over the Aegean. Note that Table 1 lists the coordinates of the areas involved in the construction of the employed indices in this study (see also Figs. 1 and 2). The easternmost latitudinal column of grid points includes information about their strength over the southeastern

TABLE 1. Acronyms and coordinates of areas delineated in Figs. 1 and 2.

Abbreviation	Explanation	Lat range	Lon range
AEG	Aegean	35°–41°N	23.5°–28.5°E
NAEG	Northern Aegean	38°–41°N	23.5°–28.5°E
SAEG	Southern Aegean	35°–38°N	23.5°–28.5°E
sEM	Small eastern Mediterranean	31°–35°N	22°–33°E
sWM	Small western Mediterranean	37°–44°N	3°–8°E
eEM	Extended eastern Mediterranean	30°–42°N	20°–35°E
eWM	Extended western Mediterranean	30°–42°N	0°–15°E

Aegean Sea and the Sea of Marmara. Six-hourly u and v wind components are used for the calculation of the 6-hourly magnitude and direction of the wind vector at each gridpoint. Following the principles of circular statistics that apply for circular variables [Eq. (2.2.4) in Mardia and Jupp (2000)], such as the wind direction that breaks at 360°, the daily mean or preferred wind direction at a grid point is calculated. The daily mean wind speed at each grid point is given by the arithmetic mean of the 6-hourly wind speeds. Similarly, the wind speed and direction are spatially “processed” over the Aegean region to obtain a metric of the daily intensity and preferred direction of the winds over the whole Archipelago. It can be argued that a more traditional approach could be pursued based on wind observations from one or more stations throughout the Aegean. However, these stations are unevenly distributed throughout the

Archipelago and individual observations can be influenced by the complex topography (section 1). This fine-scale information is filtered out in the gridded data that capture the synoptic-scale aspects of the circulation, which suits our intention to focus on the synoptic-scale dynamics involved. Also, the choice of gridded data allows the full representation of the structure of Etesians both horizontally and vertically, which could not be accomplished by the use of station data.

Figure 3 illustrates the day-to-day evolution (gray lines) of the ERA-40 1000-hPa wind speed (Fig. 3a) and direction (Fig. 3b) of the Etesians calculated for the randomly chosen year 1993. The seasonal cycles (black lines) represent multiyear averages calculated for each calendar day. Evidently, around mid-May there is a regime shift in the circulation over the Aegean. Compared to the winter season, which is characterized by relatively

FIG. 3. Seasonal pace (black curves) and daily evolution for the year 1993 (gray curves) of daily 1000-hPa ERA-40 (a) wind speed (m s^{-1}) and (b) direction averaged over AEG. The black dashed line in (a) depicts the wind-speed climatology whereas the black solid line shows the climatology only for days with northerly flow over the Aegean for the period 15 May–15 Oct.

strong variability in wind intensity and direction associated with frequent synoptic disturbances, during the transition toward summer, the wind regime over the Aegean shows reduced variability. This is manifested by a decrease of intense wind episodes, reduced departure from the northerly wind direction, as well as less frequent circulation shifts owing to the more infrequent passage of synoptic systems, resulting from the weakening and northward retreat of the midlatitude baroclinic zone. Later, during July–August (JA) and in agreement with Fig. 2a, the wind direction becomes north-northwest, while near the end of the period, there is a tendency for the flow to acquire a weaker westerly component. Most importantly, the steady decrease in wind speed observed throughout spring is reversed, leading to a prominent summer peak associated with the Etesians.

The Etesians are characterized by high variability with outbreaks lasting for up to a week or longer and interrupted by calmer periods. When the Etesians weaken, the flow relaxes toward a more westerly direction and during the peak of each outbreak the wind tends to blow from the northeast. In September and even in the beginning of October, Etesian outbreaks can also occur. The waning of the Etesians in late September and early October explains the weakening tendency of wind intensity (Fig. 3a). The subsequent reversal of this tendency heralds the shift toward the winter circulation regime. The return of high speed and direction variability implies that some of the late September or early October events could be categorized as “hybrid” Etesian outbreaks, when the northerly flow that prevails after a transient system enhances the weak background circulation. Similarly, some events in June could not be strictly categorized as “pure” Etesian outbreaks.

c. Definition of a day of Etesians and Etesian outbreaks

The definition of a day of Etesians is determined with the aid of the distribution of the daily mean wind speed and direction over the Aegean (see Fig. 1 in supplemental material). Each scatterplot represents a kind of phase space on which each point value gives the daily mean 1000-hPa ERA-40 wind speed and direction averaged over the region. The cloud of points includes days of the extended summer period [June–September (JJAS)] when Etesians traditionally occur or the core summer period (JA). Fine details of the density of the distributions for various locations can be seen in Fig. 4. Evidently, the occurrence of summer days with easterly or southerly flow over the Aegean is very rare (Fig. 4a). The increasing density for higher speeds within an elongated distribution (unlike the background distribution), which reaches a maximum of $5\text{--}6\text{ m s}^{-1}$, is

a manifestation of the regime of Etesians. This feature is more pronounced in JA when the Etesians are more frequent. Its positive slope implies that as Etesians intensify they acquire a more easterly component, in agreement with Fig. 3. In accordance with Fig. 2, the strongest contribution in the easterly component comes from the northern Aegean flow, while the stronger northwesterly flow over the southern Aegean shifts the distribution toward higher wind intensities (Figs. 4b,c). The regime becomes less distinct with increasing height (Figs. 4d,e), while it is not detectable in the similar profiles of 1000-hPa wind averaged over equally sized areas (see Fig. 1) over the western and eastern Mediterranean (Figs. 4g,h). The profile over the eastern Mediterranean is perfectly symmetrical, hindering any attempt to apply objective criteria to identify days of Etesian outbreaks there. Over the western Mediterranean, the bimodality of the flow associated with alternating southerly and northerly regimes can only be compared to the winter distribution over the Aegean (Figs. 4f,h).

A day of Etesians is defined as a day during which the daily mean 1000-hPa wind intensity exceeds the median wind speed (4.7 m s^{-1}) during the period 15 May–15 October, which is calculated including only days with preferred northerlies over the Aegean. The choice of this period was determined by the clear signature of the Etesians, which is first identified in mid-May and diminishes in early October (Fig. 3). Thus, any deviation from climatology and amplification of the northerly flow should be compared against the whole period when Etesians can be observed. Northern sector winds are defined by imposing the direction boundaries 45°W – 45°E marked in Fig. 4 by the two vertical lines. Most importantly, the employment of the direction limits is successful in recognizing approximately the western point of the slope of the elongated formation, while the wind-speed threshold locates the lower boundary of the core of the distribution associated with the regime of Etesians (Fig. 4a). Finally, an Etesian outbreak is defined as a series of days of Etesians exceeding the duration threshold of 1 day. Thus, weak and short-lived events, which probably cover parts of the Aegean Sea, are excluded from our analysis. In this sense, some “hybrid” Etesian outbreaks, occurring mainly in June and September (section 3b), are filtered out.

The threshold value derived with the aid of wind data at 1000 hPa should not be directly compared to similar thresholds proposed by previous studies that are acquired with observations at 10 m. Moreover, given the climatological pattern of the Etesians (Fig. 2a), these station thresholds are expected to vary for different stations over the Aegean, thus complicating any attempt to identify periods of enhanced Etesians over the whole

FIG. 4. Density (%) distribution of daily mean wind, averaged over the different domains and periods. The coordinates of the domains are given in Table 1, while these areas are delineated in Figs. 1 and 2. The distributions are based on the equivalent scatterplots (see text). A grid is fitted on each scatterplot with dimensions $10^\circ \times 0.2 \text{ m s}^{-1}$. The color-bar values represent the percentage of occurrences in the grid boxes calculated relative to the total distribution population extending over the whole phase space. The vertical dashed lines delineate the northern sector winds, while the horizontal dashed line marks the 15 May–15 Oct median of the daily mean wind speed distribution [except in (f)], calculated only for days with mean northerlies over the Aegean.

FIG. 5. (a) Incidence of ERA-40 days with Etesian outbreaks and (b) seasonal variability of Etesians' outbreak frequency during the JJAS period.

Aegean. The proposed methodology aims at the identification of uninterrupted periods during which the Etesians blow with higher-than-normal intensity over large parts of the geographically coherent region of the Aegean. Given that the threshold is determined by the distribution of the flow over the Aegean, the methodology can be applied independently for different reanalysis products or model outputs that are characterized by varying climatologies and thus thresholds.

3. Climatology of the Etesians

a. Frequency of Etesian outbreaks

With the aid of the above criteria, we distinguish the days of Etesian outbreaks. Figure 5a demonstrates the distribution of these events during the JJAS season and throughout the ERA-40 period. Consistent with the experience of operational meteorologists, the density of points appears to be higher in early August, which is confirmed by the seasonal variability of Etesian outbreak frequency depicted in Fig. 5b. The interannual variability of the monthly frequency of days associated with Etesian outbreaks stratified for individual months is depicted in Fig. 6. Further insight into the interannual variability and trends of the Etesian outbreak

frequency is obtained with the help of Table 2. All trends correspond to the Sen slope (Sen 1968), which are calculated using the nonparametric Mann-Kendall test for both datasets. Linear trends are also calculated and they are found to be similar to the Sen slope (not shown). The overall trend for June is negligible, while Etesians become less frequent during the period July–September. This signal was found to be more robust and statistically significant at 95% level for the entire JJAS period (-2.47% decade $^{-1}$) and for September (-6.35% decade $^{-1}$), implying a striking waning of the phenomenon during the end of the period (i.e., September), while there is evidence for less frequent Etesians even during the core JA period.

The negative trend for JJAS events is in agreement with the results presented by Poupkou et al. (2011) and was attributed to the decreasing trend in the pressure gradient between the central and southern Europe high pressure center and the Middle East low-pressure center, primarily caused by the weakening of the former center. However, the connection with the South Asian monsoon should not be disregarded, since recent studies revealed marked long-term trends in its seasonality featuring an earlier onset over the Arabian Sea and the Bay of Bengal in May (Kajikawa et al. 2012; Xiang and

FIG. 6. Interannual variability of monthly Etesian outbreak frequency in (a) June, (b) July, (c) August, and (d) September constructed with the aid of the ERA-40 dataset.

Wang 2013) and a reversed trend later in June. The advanced onset of the monsoon during early summer is attributed to the stronger-than-normal heat contrast between the Asian landmass and tropical Indian Ocean with one plausible factor for the warming trend being the increased premonsoon dust-aerosol loading along the Himalaya–Tibetan Plateau (Kajikawa et al. 2012). The early monsoon onset could result in the early establishment of the Etesians over the EM and this interesting connection should be further investigated with the aid of the ERA-Interim data that is constantly updated until present.

The frequency distribution of all Etesian outbreaks featuring onsets during JJAS with respect to their duration is illustrated in Fig. 7. For durations up to 4 days the population decreases steadily for longer events. If the population decrease rate had remained unchanged beyond the threshold of 4 days and throughout the duration spectrum, essentially no events could have been identified with durations beyond 5–6 days, which is the upper limit of synoptic time scales. Such a distribution is typical for phenomena and variability related to transient systems. However, the distribution slope is not uniform throughout the distribution, but within the interval of 4–8 days, a “shoulder” can be identified. This is a manifestation of the distinct dynamics governing the Etesians and results in their persistence beyond the typical synoptic time scale (see section 5). Typically, most Etesian outbreaks have durations less than 7–8 days, but a subset lasts even longer than 10 days, especially during JA when Etesian outbreaks occur most frequently. The subdivision of the distribution for JA, June, and September suggests that the aforementioned shoulder

in the distribution is more evident in JA followed by September, while its signature weakens in June. This is not surprising since in JA we expect the characteristic signature of the Etesians to dominate more effectively the frequency profile compared to the more transitional months of June and September.

b. Diurnal cycle of Etesians

Operational forecasters are familiar with the daytime intensification of the Etesians over the Aegean. Here, we reconstruct the diurnal cycle of the phenomenon at the temporal resolution of 6 h by producing the sequence of maps of ERA-40 MSLP and 1000-hPa wind averaged at the main synoptic hours (Fig. 8), only for days featuring Etesian outbreaks. Note that 3 h must be added to derive the local time over the region. During the night (0000 UTC) the Etesians acquire their diurnal minimum over the central and southern Aegean Sea. From the early morning hours (0600 UTC), they strengthen

TABLE 2. ERA-40 trends of Etesian outbreak frequency ($\% \text{ decade}^{-1}$) corresponding to month/season. Trends correspond to the Sen slope and are calculated by applying the Mann–Kendall test. Values in boldface denote trends that are significant at the 95% level.

Period	ERA-40
Jun	0.0
Jul	−1.47
Aug	−2.02
Sep	−6.35
JA	−1.82
JJAS	−2.47

FIG. 7. Frequency distribution of Etesian outbreaks with onset in JJAS with respect to their duration based on ERA-40 data. Etesian outbreaks are stratified with respect to event onset in JA, June, and September. Frequencies are calculated with respect to the total event population for the corresponding month or season.

especially over the central parts of the Archipelago, reaching their maximum after noon (1200 UTC) while they weaken after sunset (1800 UTC). The diurnal wind-speed range can exceed 2 m s^{-1} over the Cyclades

(36° – 38°N , 24° – 26°E), as seen in the difference between 1200 and 0000 UTC MSLP/1000-hPa wind in Fig. 9a. This daytime strengthening is the result of the increased pressure gradient especially over and to the east of the Cyclades,

FIG. 8. Diurnal evolution of the Etesians during JA. ERA-40 1000-hPa wind direction (arrows), wind speed (m s^{-1} ; shaded), and MSLP (hPa; dotted lines) averaged at the main synoptic times (a) 0000, (b) 0600, (c) 1200, and (d) 1800 UTC during Etesian outbreaks. Local summertime over the Aegean is UTC + 3 h.

(a) 12 - 00 UTC MSLP, 1000 hPa WIND **(b) DAY - NIGHT TPTT**

FIG. 9. (a) Difference between daytime (1200 UTC) and nighttime (0000 UTC) 1000-hPa wind speed (color coded), direction (arrows), and MSLP (contours). Solid (dashed) contours correspond to positive (negative) MSLP difference with a 0.3-hPa contour interval. (b) Difference between mean daytime (0600–1200 UTC) and nighttime (1800–0600 UTC) lower-troposphere diabatic heating rate TPTT (K h^{-1}), calculated by averaging TPTT at 1000, 925, and 850 hPa. Arrows and contours depict the daytime minus nighttime wind and MSLP difference as in (a). All composites are calculated for JA days featuring Etesian outbreaks.

which in turn occurs in response to the daytime strengthening of the near-surface thermal low over Turkey. After sunrise, the surface heating over the Anatolian Peninsula and the fast warming of the overlying tropospheric layers

is evident in the difference between the low-level diabatic heating rate averaged during the daytime period (0600–1800 UTC) and that during the night (1800–0600 UTC). The low-level diabatic heating rate is obtained by

averaging TPTT for the pressure levels 1000, 925, and 850 hPa. This warming is stronger over the areas of highest negative pressure anomalies, highlighting the contribution of surface warming to the intensification of the thermal low.

The lack of other significant diabatic heating sources, such as latent heat (Tyrlis et al. 2013), and the negligible diurnal variability of diabatic heating over the surrounding water masses imply that the diabatic heating during the day is largely caused by sensible heating due to increased insolation. Local airmass redistributions (e.g., sea–land breezes) result in higher surface pressure over the colder Mediterranean Sea during daytime, enhancing the pressure gradient and winds over the central and southern Aegean. Note that the diabatic heating is partly offset by advective cooling especially over western Turkey, which is exposed to the Etesians (Tyrlis et al. 2013). Thus, the actual diurnal cycle of the 2-m temperature is much less compared to the heating rates depicted in Fig. 9b.

Although data with higher resolution could reveal fine details of the daily variability of the Etesians and their interplay with the local sea–land circulations over the complex topography of the region, it is possible to interpret the salient features of the patterns depicted in Fig. 9 with the aid of Figs. 8a and 8c. The narrow band of higher daytime pressure, which extends northward off the coast of continental Greece, is combined with the lower daytime pressure over southwestern Turkey leading to a stronger daytime pressure gradient and Etesians to the east of 24°–25°E over the central and southern Aegean. Thus, the direction of the pressure gradient does not change but it enhances during the day, which explains the anomalous northerly flow depicted in Fig. 9. A similar daytime intensification of the northerly flow occurs along the coasts of the Black Sea and northwestern Turkey. Very interestingly, a reversed diurnal cycle of the Etesians intensity is observed over the northern Aegean and to the west of Cyprus. The daytime lowering of the pressure over the Balkans results in a weakening of the northwest–southeast pressure gradient over the northern Aegean, causing a weakening of the Etesians over the region. This tendency is stronger over southern Turkey because of additional changes in the direction of the pressure gradient caused by the diurnal oscillation between the lower nighttime pressure over the warm waters of the Mediterranean and the lower daytime pressure over land in Turkey (Fig. 8).

This is not the case to the east of Cyprus, where the climatological flow is southwesterly and the daytime strengthening of the thermal low over the land in the Middle East results in a stronger southwesterly flow. It is also of interest that the northerly flow weakens slightly

during the day over the Ionian Sea, especially over its northern parts (Fig. 8), although a similar climatological pressure gradient is found there as over the southern Aegean. However, here the diurnal changes in the direction of the pressure gradient are different. Unlike the southern Aegean region, the main landmass and thus daytime surface diabatic heating lies to the north or northeast of the region, especially over the Balkans. This is more evident for the areas to the south of the Peloponnese and the northern Ionian Sea and it could contribute to a daytime deceleration of the northerly flow. The east–west pressure gradient and northerly flow are less reduced only around 35°N, 18°–19°E.

4. Dynamical harbingers of Etesian outbreaks

To explore the dynamics leading to the high frequency variability of Etesians (Fig. 3), we identified all onset days of Etesian outbreaks over the Aegean during the JJAS period from the ERA-40 dataset. Figure 10 presents the evolution of MSLP and 500-hPa geopotential height fields (contours), as well as their anomalies (color shades) constructed by averaging these fields for a selection of days prior to and after the onset day of events occurring in JA. Left (right)-tilted hatching marks the areas where the positive (negative) anomalies are found to be statistically significant at the 95% level following Brown and Hall (1999). A wave disturbance originating over the North Atlantic up to 4 days prior to the onset of the Etesians can be identified. This composite signature of the surface and midtropospheric anomalies is characterized by a slight westward tilt with height. In its southeastward progression, it causes rapid circulation changes and dynamical responses over the Aegean throughout the troposphere. Some aspects are highlighted in Fig. 11, which depicts pressure–latitude vertical cross sections showing the anomalous composite signature (color coded) of all wind components (Figs. 11a–f) and potential vorticity (PV; Figs. 11g,h) averaged over the longitude band 24°–28°E and corresponding to 1 day prior to and 1 day after the onset of Etesian outbreaks occurring in JA. The 2 potential vorticity unit (PVU) surface (thick line), which corresponds to the dynamical tropopause, and isentropes (potential temperature isolines) are also depicted. Further insight into the horizontal distribution of PV at midlevels and its evolution is obtained through the examination of 500-hPa PV corresponding to composites of days prior to and after the onset of JA Etesian outbreaks (Fig. 12, left column). Similarly, the evolution of the 850-hPa temperature and wind (Fig. 12, right column) illustrates the changes in the circulation and airmass advection associated with the approaching wave

FIG. 10. Composites of (left) anomalous MSLP and (right) 500-hPa geopotential height corresponding to (a),(b) 3 days and (c),(d) 1 day prior to onset, as well as (e),(f) 1 day and (g),(h) 3 days after onset of Aegean Etesian outbreaks starting in JA. Anomalies (color coded) are calculated with respect to JA climate, while contours depict the evolution of the actual fields. Left (right)-tilted hatching marks areas where the positive (negative) anomalies are statistically significant at the 95% level.

train. In most cases the 850-hPa level is located directly above the boundary layer and is useful for tracing air masses.

Around 3–4 days prior to the onset of an Etesian outbreak, anomalous low-level southwesterly flow prevails over the central Mediterranean causing advection of warm air originating in northern Africa (Fig. 12b),

ahead of the incoming trough (Figs. 10a,b). As it moves eastward, its signature is evident over the Aegean 1–2 days prior to the onset day and extends farther aloft, in association with a relatively strong southwesterly subtropical jet (Figs. 11a,c). The induced relaxation of the pressure gradient over the EM results in the weakening of the Etesians and periods of warm weather and occasional

FIG. 11. Changes in the vertical structure of ERA-40 circulation anomalies of (a),(b) meridional and (c),(d) zonal wind, as well as (e),(f) vertical velocity ($-\text{Pa min}^{-1}$) and (g),(h) PV averaged over the longitude band 24° – 28°E (Aegean Sea). The vertical profiles correspond to (left) 1 day prior to and (right) 1 day after onset of Etesian outbreaks in JA. All anomalies are calculated with respect to the JA climate. Red (blue) values denote an intensification (weakening) in the southerly/westerly flow, ascending motion, and a stratospheric intrusion. Solid lines illustrate the JA climatology of isentropes. The lower gray thick line marks the mean JA position of the dynamical tropopause (2 PVU). The lower black thick line marks the instantaneous position of the tropopause as the phenomenon evolves.

heat waves over the region. The hot weather is abruptly terminated because of the strong cold-air advection (Figs. 12d,f) at the eastern flanks of the approaching ridge. The cold-air advection is triggered by the development of a strong ridge over central Europe that decelerates its advance as it approaches the Balkans (Figs. 10c–f). The onset of an Etesian outbreak is signaled by the rapid

transition from an anomalously southwesterly to a stronger-than-normal north-northeasterly flow at all levels (Figs. 11a–d and 12f,h). Over the EM and especially over the Aegean, strong subsidence occurs during summer, as air of midlatitude origin subsides in the vicinity of sharply sloping isentropes in its southward movement (Tyrlis et al. 2013). The weakening of the

FIG. 12. Evolution of anomalous signature (color coded) of (left) 500-hPa PV and (right) 850-hPa wind and temperature corresponding to (a),(b) 3 days and (c),(d) 1 day prior to the onset, as well as (e),(f) during the onset day and (g),(h) 1 day after onset of Etesian outbreaks occurring in JA. All anomalies are calculated with respect to the JA climate. Contours depict the instantaneous composite fields for the aforementioned selection of days with intervals of 0.1 PVU and 4°C . Note that color bar values for PV are multiplied by 0.01. For illustration, wind vectors are plotted every third column or row of grid points.

northerly flow due to the anomalous southerly flow component 1 day prior to the onset causes weaker subsidence or even upgliding of the air masses following the isentropes within the anomalous southwesterly flow. On the other hand, 1 day after the onset stronger-than-normal subsidence is observed especially over the Aegean and farther north because of the stronger northerly

flow associated with the Etesians (Figs. 11e,f). Note the low-level hot spot of vertical velocity anomaly over the Aegean corresponding to an increase in subsidence exceeding 3 Pa min^{-1} .

The arrival of the Etesians is accompanied by profound changes in the depth of the troposphere over the region that are reflected in the morphology of the

tropopause in the form of ripples on the dynamical tropopause or equivalently in the distribution of PV at 200 hPa (not shown). The aforementioned wave disturbance is also manifested farther below in the mid-troposphere, where higher (lower) 500-hPa PV values are observed in the vicinity of the trough (ridge), as depicted in Fig. 12. A broad area of tropopause depression approaches the Aegean 1 day prior to the onset as the trough moves southeastward (Figs. 10c,d). Around the time that the high-PV air of stratospheric origin reaches the sloping isentropes over the EM, it subsides and favors the development of tropopause folds (Fig. 11h). Since diabatic processes are negligible over the region in summertime upper troposphere, PV is conserved (Holton et al. 1995) as the air descends, and this explains the amplification of the positive PV anomaly in the midtroposphere over the region, which subsequently becomes stretched and elongated in the meridional direction (Figs. 12c,e,g). Note that the tropopause folds are more clearly visible in individual case studies (see Fig. II in supplemental material).

Around 3–4 days after the onset, the high-PV region may continue its eastward journey following the wave activity toward central Russia or toward the Middle East where it gradually weakens as it enters an area dominated by the baroclinic structure induced by the South Asian monsoon. Note that the tendency for wave activity to move toward Asia is more evident near the surface while at midlevels the ridge tends to spread throughout Europe (Figs. 10g,h) and a new positive height anomaly appears over western Europe. This can be explained by the fact that the duration of an outbreak could be variable and it can be followed by a new incoming wave disturbance leading to a new event.

The lack of statistically significant anomalies over the Middle East that could be related to the amplification of the Persian trough or the extension of the thermal low in the vicinity of Cyprus preceding outbreaks implies that the Etesian outbreaks are controlled by midlatitude dynamics (Fig. 10). The South Asian monsoon may control the seasonal cycle and arrival of the Etesians or contribute to their interannual variability by building the thermal low over the Middle East (Tyrlis et al. 2013), but evidently the midlatitude dynamics influence (on shorter time scales) the intensity of the western center of action that contributes to the pressure gradient over the EM (i.e., the ridge over central Europe and the Balkans). This ridge is commonly considered as an expansion of the Azores anticyclone toward Europe, though the analysis by Prezerakos (1984) indicated that this is not strictly accurate as in most cases a trough separates the Azores and the central European anticyclones, suggesting the existence of two isolated systems.

In agreement with that study, the inspection of the 500-hPa geopotential height fields clearly shows a trough following the ridge stretching in the southwesterly–northeasterly direction toward central Europe (Figs. 10d,f). In fact, the signature of the Azores anticyclone diminishes quickly from the surface upward and is very weak at 500 hPa, unlike the dominant ridge over Europe. This fact alone suggests that other dynamics that are not related to a mere intensification or displacement of the Azores anticyclone are responsible for the latter ridge, which appears to be an extension of the anticyclone that dominates the midlevel circulation over northwestern Africa (see also Zarrin et al. 2010). As the wave disturbance approaches western Europe, upstream of the trough there is poleward transfer of anticyclonic vorticity from the anticyclonic center over northwestern Africa that results in the further amplification of the preceding ridge and its evolution into a dominant circulation feature.

Thus, the timing of the Etesian outbreak onset and decay is influenced by the midlatitude dynamics, leading to successive disturbances sustaining the Etesians. Depending on the synoptic conditions, the continuous reinforcement of the ridge by new midlatitude disturbances can lead to rather prolonged Etesian outbreaks. Also, teleconnection patterns that influence the intensity and paths of the midlatitude synoptic systems, such as the North Atlantic Oscillation, can in turn influence the frequency of the Etesians on intraseasonal or interannual time scales. Chronis et al. (2011) presented evidence that during the positive phase of the North Atlantic Oscillation, the summer meridional wind magnitudes (Etesians) are enhanced over the EM. Most importantly, the results described above are robust and not sensitive to the exact choice of the thresholds used for the definition of a day of Etesians over the Aegean. Even for significantly lower or higher thresholds, the signature of the midlatitude influence remains robust with only varying amplitudes and phase shifts of the anomalies owing to the temporal changes in the definition of the onset days.

5. The persistence of the Balkans anticyclone and Etesians

Investigation of the evolution of the horizontal arrangement of the thermodynamic equation terms in the middle troposphere sheds further light on the dynamical processes occurring before and after the Etesian outbreaks (Fig. 13). Ascent and consequent cooling due to vertical advection are observed slightly ahead and within the trough (Figs. 10 and 13). The anomalous cooling temporarily perturbs and weakens the adiabatic warming owing to subsidence that dominates over the area, featuring a maximum over Crete. Simultaneously with

FIG. 13. Evolution of ERA-40 500-hPa (a),(b) VATT, (c),(d) HATT, and (e),(f) TPTT for composites of days corresponding to (left) 3 days prior to and (right) onset day of Etesian outbreaks occurring in JA. Shaded values correspond to anomalous fields with respect to the JA climate while contours show the instantaneous composite fields. Contour interval is 0.5. Solid (dashed) contours correspond to positive (negative) values. Contours for -0.5 , 0 , and 0.5 K day^{-1} are suppressed.

the surface warming (Fig. 12b), weak warm-air advection occurs aloft and ahead the approaching trough, as the southwesterly flow strengthens (Fig. 13c). The positive TPTT anomaly accompanying the trough (Figs. 13e,f) can be related to unsettled weather and convective activity leading to latent heat release.

After the passage of the trough, the southward advance of air masses along the sharply downward-sloping

isentropes results in a positive VATT anomaly owing to the enhanced subsidence. The excess warming is counterbalanced by anomalous cold advection as the Etesians build up (Figs. 13b,d). Unlike the negative HATT anomaly, the positive VATT anomaly extends farther to the north closer to the core of the ridge, where diabatic cooling becomes the second most important term in the thermodynamic equation. Since anticyclones

produce fair weather, the overwhelming proportion of the diabatic cooling (Fig. 13f) in their vicinity refers to radiative cooling under the clear skies. This corroborates the diabatic enhancement hypothesis proposed by Rodwell and Hoskins (1996, 2001), indicating that the weaker background subsidence can be further enhanced since subsidence results in adiabatic warming, a reduction in the specific humidity of air parcels, a lowering of the radiative level with increased radiative cooling in the cloud-free atmosphere, and, therefore, diabatic descent. The air parcels descend in parallel but cut across the sloping isentropes and their potential temperatures decrease (i.e., cooling occurs; see also Rodwell and Hoskins 1996). A suitable environment for this to occur is the ridge of the incoming wave disturbance.

Apart from the contribution of the anticyclonic vorticity transfer in the amplification of the ridge over central Europe, described in section 4, which is also hinted by the expansion of the negative PV anomaly over northwestern Africa (Figs. 12e,g), the ridge and surface anticyclone may also be reinforced because of the diabatic enhancement mechanism. As the ridge approaches southeastern Europe, where background subsidence dominates, it triggers further subsidence through a positive feedback loop. Subsidence is enhanced in the free troposphere under the clear skies and this in turn enhances the anticyclonic circulation below and also the subsidence–cold advection–northerly flow at the eastern flanks of the system.

Although a more detailed investigation of individual cases could be more revealing compared to composite signatures, some aspects of the evolution of the aforementioned anomalies can be explained with the aid of the Rossby wave theory. Over regions of strong background westerly flow, such as over the Atlantic, short-wave Rossby waves are generally not characterized by sufficiently high westward phase speed to overcome the background eastward flow and thus they propagate eastward. However, as they move toward southeastern Europe where the background flow weakens, the wave disturbances are expected to slow and this allows other processes, such as the diabatic enhancement mechanism to interact more efficiently with them. This influence may lead to an increase in the size or wavelength of the disturbance, which is actually hinted in Fig. 10f by the expanding ridge, and further deceleration. Eventually, the amplifying ridge appears to stagnate over the Balkans, which explains the persistence of the Etesians beyond synoptic time scales and over periods of several days (Fig. 7). This feedback may be interrupted when a new trough approaches from western Europe, resulting in a temporary breakdown of the Etesians.

6. Summer tropopause folds over the Aegean

As discussed in the previous section, the lower tropopause and positive PV anomaly over the Aegean observed with the onset of Etesian outbreaks (Fig. 11h) indicate the development of tropopause folds (e.g., Appenzeller and Davies 1992; Thorpe 1997; Sprenger et al. 2003). Inspection of the evolution of the vertical structure of the troposphere before and after the onset of numerous onsets of Etesian outbreaks, such as the one depicted in Fig. II in the supplemental material, reveals that the Aegean Sea marks the longitude at which the tropopause folds form and progress downward and southeastward, allowing high-PV air of stratospheric origin to reach as low as 700 hPa. Note that in the case studies presented in this study, tropopause folds are identified by inspection of the topography of the dynamical tropopause (2-PVU surface). A fold occurs in areas where multiple crossings of the 2-PVU surface are identified in the vertical direction. Folds preferentially form over the Aegean where the sharp meridional and zonal slope of the isentropes, which is induced by the South Asian monsoon, allows interaction between the troposphere and the stratosphere (Figs. 11g,h) and also the abrupt sharpening of the folds. This air is trapped in the northwesterly midlevel flow and low-level jets that transfer air originating in the middle troposphere over the EM toward the Red Sea and the Persian Gulf region (Tyrlis et al. 2013).

The occurrence of tropopause folds over the Aegean is so frequent and widespread that it results in a spatially locked and semipermanent phenomenon during the summer. Its footprint can be easily detected by comparing the seasonal cycle of PV in the upper and middle troposphere averaged over two equally sized regions over the eastern and western Mediterranean and specifically over the Aegean (Figs. 14a,b). Details of the seasonal cycle throughout the depth of the troposphere are illustrated in Fig. 14c and in Fig. III in the supplemental material, both of which depict meridional and zonal vertical cross sections of the difference of PV and the position of the dynamical tropopause between July and January averaged over the latitude and longitude extent of the Aegean sector (36° – 40° N, 24° – 28° E). In agreement with the weakening of the mean pole-to-equator temperature gradient in the Northern Hemisphere during boreal summer and the associated weakening and poleward shift of the jet stream, which develops below the tropopause, the tropopause over the midlatitudes tends to be located higher compared to during winter. This is indicated by the significant decrease in PV values within the 300–100-hPa layer over the Aegean sector (30° – 40° N; Fig. 14c) and by the summer upper-level PV

FIG. 14. (a) Seasonal evolution of PV averaged over the upper troposphere (300–200 hPa) across AEG (green curve: 34°–41°N, 23.5°–28.5°E), the eastern Mediterranean (eEM) (red curve: 30°–42°N, 20°–35°E), and the western Mediterranean (eWM) (blue curve: 30°–42°N, 0°–15°E). AEG, eEM, and eWM are delineated in Figs. 1 and 2 and their coordinates are listed in Table 1. (b) As in (a), except the seasonal cycle of PV is averaged over the midtroposphere (600–300 hPa). All seasonal cycles are smoothed with an 11-day running average to produce less noisy profiles. (c) Meridional pressure–latitude vertical cross section depicting the difference July minus January PV averaged over the 24°–28°E sector (PVU; color fill) and July climate isentropes (contours). Gray (black) thick line marks the position of the 2-PVU dynamical tropopause during January (July). Orography is superimposed.

minimum over all areas (Fig. 14a). The tropopause is elevated especially over Asia under the influence of the South Asian monsoon. However, in the midtroposphere the seasonal cycle features a prominent summer peak over the EM and especially over the Aegean (Fig. 14b). Unlike the winter maximum characterized by high variability due to transient tropopause folds, the more stable summer maximum is associated with

the quasi-permanent nature of the folds, allowing intrusions of high PV air. The profiles for the western Mediterranean and the EM start to diverge during late May and converge in early October, signifying the influence of the South Asia monsoon over the EM during exactly this period. In fact, the EM profile is synchronous with that of the Etesians' intensity (Fig. 3), their frequency (Fig. 5b), and even the subsidence over the region (Tyrllis et al. 2013).

During the maximum tropopause folding activity and Etesian outbreaks frequency in late July and early August, the tropopause at the latitude of the Aegean is in fact located at a slightly lower pressure level compared to winter, marking a narrow corridor around 400–300 hPa of positive PV anomalies highlighting the more frequent subsidence of high PV, in contrast to neighboring regions (Fig. 14c). A second “transport corridor” is evident over central Asia that needs to be studied further (see Fig. III in the supplemental material). Detailed inspection of cross sections at various latitude–longitude bands shows that the Aegean transport corridor is distinct and independent from the mid-/low-level elongated band of positive PV anomalies whose altitude decreases with latitude and whose top is located at around 500 hPa over the Aegean. It lies above a negative PV anomaly. Latent heating sources can induce a dipole of PV anomalies with PV destruction (generation) above (below) the diabatic heating sources in the middle (lower) troposphere (Hoskins et al. 1985; Hoskins 1991). The fact that the latent heating sources diminish during the summer in the Northern Hemisphere and in combination with the tendency of the altitude of the convection/clouds to reduce with latitude could possibly explain the structure and distinct character of the mid- and low-level PV anomalies depicted in Fig. 14c.

The formation of tropopause folds is common in the midlatitudes during winter and transitional seasons, as they develop in close dynamical association with transient synoptic systems. They induce stratosphere-to-troposphere transport—that is, the irreversible intrusion of dry, high-PV, and ozone-rich air (Holton et al. 1995). Although wintertime or springtime events of stratospheric intrusions over the EM have been the subject of numerous studies (Gerasopoulos et al. 2006; Akritidis et al. 2010), explicit reference to summer events is only made in the short 1-yr climatology of tropopause folds and cross-tropopause exchange presented by Sprenger et al. (2003). To a large extent, photochemistry can explain the observed background ozone levels in the troposphere (Crutzen et al. 1999; Lelieveld and Dentener 2000). However, the contribution of the stratospheric ozone reservoir through deep stratospheric intrusions (in the vicinity of a tropopause fold) to the formation of the mid- and low-level ozone pools observed over the EM and the Middle East (Roelofs and Lelieveld 1997; Lelieveld et al. 2009) is an intriguing subject of ongoing research. The EM and the Middle East, where subsidence is observed throughout the tropospheric column during summer, is evidently a region with abundant tropopause-fold occurrence and significant stratospheric contribution to the low-level ozone pool observed during summer (Zanis et al. 2013). On the other hand, the

persistent northerly flow associated with the Etesians controls significantly the transport of air masses from Europe toward the EM that are rich in ozone precursors (e.g., Kalabokas et al. 2008).

The Etesians represent the lower-level manifestation of changes occurring in the summer troposphere over the EM and are not dynamically independent to the formation of tropopause folds. Indeed, the midlevel PV maxima resulting from stratospheric intrusions tend to accompany the Etesians’ intensity maxima (see Fig. IV in the supplemental material). The connection between the tropopause-fold dynamics and the Etesians could be described with the aid of the PV– θ framework (Hoskins et al. 1985; Hoskins 1991). The development of a positive upper- and midlevel PV anomaly in the vicinity of the tropopause fold and the stratospheric intrusion is linked to a positive (negative) θ anomaly above (below) the PV anomaly and cyclonic circulation around it, while anomalous ascent (subsidence) occurs ahead (behind) the anomaly. As this anomaly moves southeastward and downward (Fig. 12), an enhancement of the subsidence and the northerly flow occurs upstream of the PV anomaly. Although further study is needed, inspection of several case studies (see Fig. II in the supplemental material) revealed that it may take around 1 day for the enhanced subsidence and northerly flow behind the anomaly to reach the lower troposphere and lead to an Etesian outbreak and the related stronger-than-normal subsidence over the Aegean, as depicted in Fig. 11. Indeed the maxima of midlevel PV and intensity of the Etesians are not always perfectly synchronous (see Fig. IV in the supplemental material). Also, depending on the vertical structure of the troposphere, a strong fold is not always connected with a strong episode of Etesians over the Aegean and vice versa.

7. Summary and conclusions

The Etesians are recurrent and persistent northerly winds that dominate the summer EM circulation. This study builds on Tyrlis et al. (2013), who suggested that the Etesians and subsidence over the region are reconciled manifestations of the South Asian monsoon. The variability of the Etesians results from processes on various space and time scales, including a distinct diurnal cycle. On intra- and interannual scales their variability is influenced by the monsoon, which essentially regulates the seasonal cycle of the background-flow regime. The shorter-term variability is controlled by the midlatitude dynamics and features alternating phases of enhanced northerly flow (Etesian outbreaks) interrupted by quiet spells. Considering the ventilating effect of the Etesians, the latter are often associated with heat waves. We

compiled a climatology of the Etesian outbreaks and explored the tropospheric dynamics governing the mid-latitude influence and the diurnal cycle of the Etesians. The methodology is based on the probability distribution of the circulation state over the Aegean, which reveals the unique signature of the regime of Etesians and allows the choice of objective criteria for its identification.

The Etesian outbreaks are most abundant from mid-July to mid-August, while a distinct maximum was found in the first week of August. In agreement with previous studies, a negative trend of the Etesian outbreak frequency is identified during the JJAS period. Stratification with respect to the month revealed that this trend is very strong in September, but it diminishes during the beginning of the period in June. The significant diurnal variation of the Etesians, previously recognized empirically, is confirmed by our analysis. The daytime deepening of the Anatolian thermal low, which is driven by near-surface sensible heating, strengthens the pressure gradient and intensifies the Etesians over the central and southern Aegean. On longer time scales, the midlatitude influence on the EM circulation arises from wave disturbances originating over the North Atlantic, identified as early as 3–4 days prior to the onset of Etesian outbreaks. They can trigger strong ridges over the Balkans that perturb the background state induced by the monsoon. The ridge amplifies because of the northward transfer of anticyclonic vorticity from northwestern Africa, augmented possibly by the diabatic enhancement mechanism, whose positive feedback can explain the deceleration and anchoring of the ridge over the Balkans. Although the midlatitude forcing takes place on synoptic time scales, this mechanism may explain the persistence of Etesian outbreaks beyond synoptic time scales. Provided that the outbreaks survive temporary breakdowns of the Balkan ridge, their revival from new wave disturbances can result in longer-lasting events.

The arrival of the ridge over the Balkans is accompanied by a sudden change in the circulation over the Aegean from southwesterly to northerly at all levels, while the cold advection sometimes ends heat-wave episodes. As air masses travel southward, they descend in the vicinity of the sharply sloping isentropes, leading to tropopause foldings and stratospheric intrusions of dry air. The background structure featuring the sloping isentropes is the result of the zonal asymmetry induced by the monsoon that ultimately favors the tropopause foldings. In fact, these events are so frequent that a quasi-permanent tropopause fold appears in the JA climate, while the seasonal cycle of the midtropospheric PV averaged over the EM has a prominent summer peak that does not occur

over the western Mediterranean—an area not influenced by the Etesians.

In conclusion, the various components observed over the EM that include the Etesians, subsidence, tropopause folds, stratospheric intrusions, and the summer ozone pool are dynamically interwoven manifestations of the influences induced by the South Asian monsoon and the midlatitudes, both of which act on various time scales and interact with the local orography. As a result, the Etesians are an integral component of the global climate and quantification with the aid of our index can facilitate the understanding of related dynamics, such as the connection with the local Hadley circulation over eastern Africa, possibly because of regulation of the moisture flux from the Mediterranean depending on the intensity of the Etesians. In the framework presented in this study, midlatitude teleconnection patterns such as the North Atlantic Oscillation can potentially introduce a climatic signal over the EM, while tropical phenomena such as ENSO can affect the climate of the region on longer time scales by influencing the monsoon.

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