The Origin of Western European Warm-Season Prefrontal Convergence Lines

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ABSTRACT

The authors investigate the origin of prefrontal, warm-season convergence lines over western Europe using the Weather Research and Forecasting Model. These lines form east of the cold front in the warm sector of an extratropical cyclone, and they are frequently the focus for convective development. It is shown that these lines are related to a low-level thermal ridge that accompanies the base of an elevated mixed layer (EML) plume generated over the Iberian Peninsula and northern Africa. Using Q-vector diagnostics, including the components that describe scalar and rotational quasigeostrophic frontogenesis, it is shown that the convergence line is associated with the rearrangement of the isentropes especially at the western periphery of the EML plume. The ascending branch of the resulting ageostrophic circulation coincides with the surface velocity convergence. The modeling results are supported by a 3-yr composite analysis of cold fronts with and without preceding convergence lines using NCEP–NCAR Reanalysis-1 data.

1. Introduction

Warm-season thunderstorms over western Europe often develop along surface wind-shift lines characterized by horizontal convergence (hereafter referred to as convergence lines) within the warm sector east of the cold front of an extratropical cyclone. Typically, these lines first appear over France or the Bay of Biscay ahead of a cold front and then spread northeastward. The convergence lines are accompanied by a trough in the surface-pressure field and sometimes a rather expansive, closed low pressure area develops in the warm sector (van Delden 1998, 2001). Although prefrontal rainbands and wind-shift lines in the warm sector of extratropical cyclones have been studied rather extensively (e.g., Browning and Monk 1982; Hobbs et al. 1990; Hobbs et al. 1996; Locatelli et al. 1997; Schultz and Steenburgh 1999; Stoelinga et al. 2003; Koch 2001; Schultz 2005), the dynamical origin of warm-season western European convergence lines has barely been addressed in the formal literature despite their ubiquity. Van Delden (1998, 2001) linked the formation of the convergence line (in his studies referred to as a thundery low) to the intensification of the cold front as warm air is advected northward from the Iberian Peninsula, while Kaltenböck (2004) proposed that the origin of these lines is tied to differential diabatic heating. Herein, we use the Weather Research and Forecasting Model and apply Q-vector diagnostics, relating the convergence lines to the presence of an elevated mixed layer. The results are put into a broader context by a 3-yr composite analysis.

Some background on Q-vector diagnostics and elevated mixed layers is provided in the next section. The modeling framework will be introduced in section 3. We will then present the modeling results as well as a composite analysis in section 4, and a discussion is provided in section 5. Finally, a summary and a conceptualization of the results are offered in section 6.

2. Background

a. The Q vector and its decomposition

To gain a basic understanding of the origin of prefrontal convergence lines, quasigeostrophic (QG) theory will be applied. Although at the scale of the line non-QG effects may be relevant, we merely seek a qualitative understanding of what flow features are associated with forcing
for low-level upward motion and hence do not require a quantitative solution for upward velocity [qualitatively, the forcing for vertical motion and the vertical motion field itself as diagnosed from QG principles are sufficiently similar to higher-order approximations; see Bosart and Lin (1984), their Fig. 3]. With that in mind, we only expect to find broad signals in our analyses, rather than an exact reflection of the convergence lines. Herein, we use the Q vector, which elegantly quantifies the instantaneous rate at which thermal wind balance is destroyed by the geostrophic flow (Hoskins et al. 1978; Davies-Jones 1991; Bluestein 1993, 360–361; Davies-Jones 2009). This disruption of thermal wind balance may be described in terms of the time rate of change of the horizontal (potential) temperature gradient (i.e., vector frontogenesis) following the geostrophic motion (Hoskins et al. 1978). In other words, forcing for QG vertical motion is related to the reconfiguration of isotherms (or isentropes) by the geostrophic wind if diabatic effects are neglected. The Q vector conveniently points in the direction of the lower branch of the resulting ageostrophic circulation and thus directly indicates where that circulation has a vertical component (Hoskins and Pedder 1980). Here, we use the formulation of the QG omega equation put forth by Keyser et al. (1992) and Martin (1999a,b), who explicitly considered projections of the Q vector tangential and normal to the local horizontal (potential) temperature gradient:

\[
\left( \sigma \nabla_p^2 + f_0^2 \frac{\partial^2}{\partial p^2} \right) \omega = -2 \nabla_p \cdot Q_n - 2 \nabla_p \cdot Q_s, \tag{1}
\]

where \(\sigma\) is the static stability parameter, \(f_0\) is the Coriolis parameter on the \(f\) plane, \(\omega\) is the vertical motion in pressure coordinates, and \(Q_n\) and \(Q_s\) are the projections of the Q vector normal and tangential to the local isentropes (or isotherms), respectively (Keyser et al. 1988, 1992; Martin 1999a,b).
The two Q-vector projections highlight regions where parcels following the geostrophic flow experience (i) traditional (scalar) frontogenesis, whereby the magnitude of the parcels’ horizontal (potential) temperature gradient changes (diagnosed by $Q_n$), and (ii) rotational frontogenesis whereby the orientation of the parcels’ horizontal (potential) temperature gradient changes [diagnosed by $Q_s$; Keyser et al. (1988, 1992); Kurz (1992); Martin (1999a,b)]. The divergences of $Q_n$ and $Q_s$ are related to transverse and shearwise vertical motion, respectively (Martin 2006). Since the motion of features in the (potential) temperature field in the lower free atmosphere is strongly influenced by advection by the geostrophic wind, the frontogenesis functions loosely describe the development of the thermal boundaries themselves [e.g., changing curvature of the isotherms or isentropes on a pressure surface being accompanied by $Q_s$ convergence (Martin 1999b)].

To relate the surface convergence lines to Q-vector forcing, the following assumptions are made. Based on the analyses by Clough et al. (1996) and Davies (2015), who discuss Green’s function solutions of the omega equation,1 the qualitative solution of Eq. (1) corresponds to the negative of the smoothed version of the forcing function, with the condition that the vertical motion vanishes at the ground (assuming a level surface).2 For instance, a region of negative Q-vector divergence in the lowest kilometers of the atmosphere will be accompanied by upward motion that is maximized where the forcing function has a minimum (cf. Davies 2015, his Fig. 1). The link to the convergence lines is established by observing that a linearly aligned regime of low-level rising motion is associated with a linear region of horizontal wind convergence at the surface via mass continuity. In the QG picture, vertical stretching of planetary vorticity within the upward motion regime will lead to an increase in cyclonic geostrophic vorticity and hence falling surface pressure, consistent with these convergence lines being accompanied by reduced pressure.

In short, the quasi-geostrophically forced part of the prefrontal convergence must lie beneath or within a region dominated by Q-vector convergence. With this in mind, we can assess whether the formation of the convergence lines can be explained with QG principles, and more importantly, which features of the synoptic-scale flow lead to the relevant Q-vector configuration. Before applying this concept, we present a brief review of elevated mixed layers.

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1 Another, more common, but in some cases perhaps less enlightening, approach is to assume the forcing functions to be sinusoids (Durran and Snellman 1987).
2 In pressure coordinates, the assumption is usually made that vertical motion vanishes at 1000 hPa (e.g., Bosart and Lin 1984).
b. Elevated mixed layers

The first step in the formation of an elevated mixed layer is the presence of a deep and dry convective boundary layer (CBL), typically developing over arid, elevated terrain during the warm season (Bennett et al. 2006). This CBL, characterized by nearly constant potential temperature and water vapor mixing ratio with height, may subsequently be advected off the plateau atop cooler air, whereupon it is referred to as an elevated mixed layer (EML; Carlson et al. 1983). Over the United States, EMLs typically originate from the elevated terrain of the southwestern part of the country as well as Mexico. Over Europe, the main source for EMLs is the Iberian Peninsula, and the respective EML is sometimes referred to as a Spanish plume (Carlson and Ludlam 1968; Bennett et al. 2006). More recently, the overall synoptic pattern involving an EML (rather than the EML itself) has been referred to as a Spanish plume (Morris 1986; Lewis and Gray 2010). Another important EML source is the northern Sahara, and the deeply mixed, dry, and often dusty layer is referred to as a Saharan air layer (SAL; Karyampudi and Carlson 1988). A deep CBL is also frequently found over Turkey, which may thus serve as a source for EML plumes as well, especially over eastern Europe and western Russia. An EML is typically manifest as a thermal ridge in the lower free atmosphere (e.g., at 850 hPa), and, if ascent is
present, altocumulus castellanus clouds can often be found at the top of the EML (Carlson and Ludlam 1968). The EML has long been known to be associated with often severe convective weather episodes over western Europe (Carlson and Ludlam 1968; van Delden 1998; Lewis and Gray 2010). This is in part related to the potential for large CAPE to develop in the presence of an EML (Carlson and Ludlam 1968).

3. Modeling framework

We used the Advanced Research version of the Weather Research and Forecasting Model (ARW; Skamarock et al. 2008), version 3.5.1, with a model domain covering the western and central parts of Europe (extending between 36°–62°N and 24°W–34°E), using a horizontal grid spacing of 30 km. The simulations employed vertically stretched sigma coordinates across 35 levels. Vertical grid spacing increases from about 50 m near the ground to about 800 m toward the domain top, which is at 50 hPa. The integration time step was 180 s and boundary conditions were provided by the National Centers for Environmental Prediction’s (NCEP) Global Forecast System (GFS). We used the Kain–Fritsch convection parameterization (Kain and Fritsch 1990) and the Yonsei University (YSU; Hong et al. 2006) boundary layer scheme. The simulations were initialized using 0000 UTC data prior to the formation of the convergence line and they were run for 36 h.

A challenge when applying QG principles to a mesoscale model is that the output fields are rather noisy, which is related to orographic circulations, gravity waves, parameterized convection, etc. In particular, the Q vectors and their divergence, as well as the velocity divergence field, are too noisy to be analyzed directly. We thus used a Δx = 60 km horizontal grid (i.e., every 60 km).
other horizontal grid point was removed from the original grid) and applied a Gaussian filter with a standard deviation of 2.5 \( \Delta x \) to the potential temperature \( \theta \) and the geostrophic wind before calculating the Q-vector fields. The horizontal wind convergence fields were likewise smoothed.

We will focus on the Q-vector convergence at the 850-hPa pressure level, where the thermal signature of the EML tends to be well pronounced, but where there is little orography extending into the pressure surface. At this level, the base of the EML is manifest as a thermal ridge.

4. Results

a. WRF simulations

In this section we will present in detail one typical case where a prefrontal convergence line developed and subsequently briefly introduce the general evolution of two additional cases. A composite analysis will be presented in the next subsection.

On 2 August 2012, a typical situation featuring a prefrontal convergence line was present over central Europe, as shown in Fig. 1. The wind shifted from southerly directions east of the convergence line to westerly directions west of the line, and there was a 1–2-K decrease in temperature toward the west across the line, while the dewpoint did not vary appreciably. Only across the northernmost segment of the convergence line were the temperature and dewpoint contrasts more pronounced, which was related to precipitation occurring west of the line (not shown).

To understand the origin of this convergence line, we will go back a day when the line first became apparent. At 0300 UTC 1 August 2012, WRF output reveals an upper-level trough over the eastern North Atlantic Ocean west of the British Isles and a deep southwesterly flow extending from southwestern into central Europe (Fig. 2). The cold front had just reached western Ireland. Reminiscent of the frontal fracture observed in some cyclones (Schultz et al. 1998), the temperature contrast across the front was rather poorly pronounced over Ireland. Reminiscent of the frontal fracture observed in some cyclones (Schultz et al. 1998), the temperature contrast across the front was rather poorly pronounced over Ireland. The base of the EML is visible as a thermal ridge extending from the southwestern Mediterranean Sea into France. The convergence line first became apparent early on 1 August 2012 over central France. Around that time, there was a conspicuous cloud band stretching across central France (Fig. 3a), which, however, moved rapidly northeastward (Fig. 3b) and had no appreciable reflection in the surface pressure field. By 1430 UTC, the convergence line was well apparent in the satellite imagery over central

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3 The output for 0300 UTC, rather than the initial analysis, was chosen to allow the model to develop mesoscale structures from the GFS background.

4 Convergence-line formation was preceded by a (thermal) low over the Iberian Peninsula, which then became elongated while spreading into France, which is a typical progression (van Delden 1998).
France after deep convection had begun to develop along the line (Fig. 3b). On the subsequent day, 2 August 2012, the cold front and the preceding convergence line had progressed into Germany, leading to the situation depicted in Fig. 1, and on 3 August 2012, the cold front had caught up with the convergence line over Poland (not shown).

Figures 4a,c show the simulated progression of the convergence line, represented by a broad region of surface velocity convergence, and the elevated mixed layer, revealed by maxima in the midlevel temperature lapse rates. The region of surface wind convergence is found primarily at the western periphery of the EML, moving from France into an area extending from western Germany into the North Sea. Figures 4b,d shows a region of Q-vector convergence extending from northern France across the British Isles (Fig. 4b), progressing northeastward, extending across Germany into the North Sea at 0000 UTC 2 August (Fig. 4d).

Figure 5 facilitates a more detailed look at the Q vectors, showing the isentrope-normal and -parallel Q-vector components at 0000 UTC 2 August 2012. Scalar (Fig. 5a) and rotational (Fig. 5b) frontogenesis are equally active and both contribute to forcing for upward motion. Convergence of $Q_n$ vectors along the isentropes highlights the elongation and narrowing in time of the thermal ridge (Martin 1999a), which in turn is accompanied by traditional frontogenesis (nonzero $Q_n$). In particular, the $Q_n$ vectors and their convergence are maximized at the western periphery of the thermal ridge.

The Q-vector convergence extends beyond the northwestward extent of the EML (Fig. 4), which is related to the narrowing warm sector during the large-scale occlusion process (Martin 1999a,b). Embedded within this configuration is the Q-vector convergence accompanying the EML plume, which is embedded in the warm sector as revealed by the unfiltered $\theta$ fields (Fig. 6). However, both regions of Q-vector convergence have

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5 Herein, the midlevel temperature lapse rates are calculated via $[T(850\text{ hPa}) - T(700\text{ hPa})]/[z(700\text{ hPa}) - z(850\text{ hPa})]$. 

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Fig. 8. (a) Temperature lapse rate between 850 and 700 hPa (K km$^{-1}$, color shaded) and Q-vector convergence (blue contour, $3 \times 10^{-18}$ m kg$^{-1}$ s$^{-1}$). (b) The 850-hPa potential temperature (K, color shaded) and manually added fronts at 1200 UTC 29 Jul 2009. (c),(d) As in (a),(b), but at 0000 UTC 30 Jul 2009.
merged in Figs. 4b,d because of the smoothing, such that the QG forcing associated with the EML no longer stands out.

Interestingly, the Q-vector convergence is displaced slightly to the northeast relative to the location of the convergence line for several output times. This offset is quite well pronounced at 0900 UTC 2 August, shown in Fig. 7a. A vertical cross section from southwest to northeast through the thermal ridge is presented in Fig. 7b, revealing a quadrupole-like horizontal velocity divergence pattern below the 650-hPa level consistent with a low-level circulation, as annotated in the figure. While the depth and sense of the circulation match what would be inferred from low-level Q-vector convergence–divergence couplet associated with the thermal ridge, the circulation is shifted slightly to the southwest. This is likely related to the region of Q-vector convergence above the 600-hPa level southwest of the convergence line, effectively “pulling” the circulation pattern southwestward (if, e.g., some constant upward velocity is subtracted from the circulation depicted in Fig. 7b, the circulation would shift to the northeast).

Two additional cases were investigated: 29 July 2009 and 1 August 2013. The Q-vector convergence in relation to the EML is shown in Figs. 8a,c for 29 July 2009. The strongest Q-vector convergence is associated with the EML plume, and a tendency for the Q-vector convergence to shift to the thermal axis as the EML narrows is apparent (e.g., Fig. 8c). The location of the cold front relative to the northward expanding EML is shown in Figs. 8b,d. Initially, there was a weak occluded front over the British Isles (Fig. 8b), which is poorly represented in the 850-hPa potential temperature field as the cyclone had aged considerably at that time. That boundary’s thermal gradient became better pronounced as it approached the EML (Fig. 8d).

The development on 1 August 2013 (Fig. 9) resembled the previous cases, except that there were two “surges” of northward-spreading EML plumes. The EML plumes were accompanied by regions of Q-vector convergence over western France and the southeastern United Kingdom at 1200 UTC 1 August 2013 (Fig. 9a). Twelve hours later (Fig. 9b), areas of Q-vector convergence at the western periphery of the EML are present over the
Gulf of Biscay and the North Sea. The Q-vector convergence along the northwest Mediterranean coast appears to be associated with the sharp eastern edge of the EML, but given the close proximity to the model orography, these signals are not trustworthy.

The 850-hPa potential temperature fields seen in Figs. 9b,d suggest that the Atlantic cold front slowly caught up with the EML plume.

Despite the at times spotty nature of the Q-vector convergence, there is a pattern that emerges from these simulations. The prefrontal convergence line appears to be accompanied by QG vector frontogenesis at the western periphery of (and at later stages amid) the EML plume as it is drawn into and deformed within the warm sector ahead of the cold front.

b. Composite analysis

To assess whether these results are consistent with the behavior of a larger number of cases, we performed a composite analysis using NCEP–NCAR Reanalysis-1 data (Kalnay et al. 1996) similar to the analysis done by Schultz et al. (2007). These reanalysis data are available on a global 2.5° × 2.5° grid at 17 pressure levels. Herein, we composited data from the 850- and 700-hPa levels. We selected three years (2012–14) and averaged the flow and thermodynamic patterns involving warm-season warm sectors with and without prefrontal convergence lines.

When designing the composite study, the goal was to avoid a seasonal bias. For instance, warm sectors without prefrontal convergence lines are observed in the winter as well as in the summer months, while warm sectors with prefrontal convergence lines are observed mainly during the summer months. Compositing warm sectors from every month of a year would thus have highlighted differences between the warm and the cold seasons, rather than between warm sectors with and without prefrontal convergence lines. We thus only analyzed those months in which fronts with and without preceding convergence lines occurred, which, in the dataset we analyzed, is between May and September. The following additional requirements had to be fulfilled for days to be included in the analysis: (i) a broad warm sector covered much of western and central Europe, (ii) the cold front was aligned roughly along the European Atlantic or North Sea coast at the analysis time, and (iii) three independent surface analyses [Met Office, German Weather Service, and the Berlin Weather Map (BWM) archive maintained by the Free University of Berlin] showed that a convergence line was (or was not) present. For the large majority of cases, the three analyses were consistent. When in doubt, slightly more weight was given to the BWM archive given the high observation density used to produce the BWM analyses.

During the 3-yr period, a total of 19 cases fulfilling the above criteria were found in each category (Table 1).

For an overview, we plotted the monthly distribution of occurrences of cold fronts without and with preceding convergence lines (Fig. 10). Cold fronts without convergence lines occur throughout the warm season, without a strong preference for a certain month. The occurrence of cold fronts accompanied by a prefrontal convergence line, however, peaks during August. This analysis is consistent with the presence of an EML in cases with convergence lines because the EML is most pronounced in the late summer months. The composite fields of the midlevel temperature lapse rates are shown in Fig. 11 to assess the presence of an EML. The prefrontal air mass in all cases in the dataset was located over northern France, Benelux, and northern Germany, per requirement ii above. While the EML footprint was smoothed during the compositing, a signal is still apparent showing weaker midlevel temperature lapse rates ahead of cold fronts without a convergence line (Fig. 11). In these cases, there is still an area of comparatively steep midlevel lapse rates over the Mediterranean regions, but this pattern does not extend into the warm sector over the northern parts of western and central Europe. Finally, the composited 850-hPa wind field is shown in Fig. 12. In cases without a convergence line (Fig. 12a), the flow is mostly westerly in the warm sector and has an accordingly small meridional component. When the cold fronts were preceded by a

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<th>Table 1. List of cases of cold fronts with and without preceding convergence lines used in the composite analysis.</th>
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<tr>
<td>With line (n = 19)</td>
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<td>0600 UTC 11 May 2012</td>
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<td>0600 UTC 18 Jun 2012</td>
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<td>0600 UTC 29 Jun 2012</td>
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<td>0000 UTC 5 Jul 2012</td>
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<td>0000 UTC 2 Aug 2012</td>
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<td>1800 UTC 10 Aug 2014</td>
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convergence line, the 850-hPa wind over western and central Europe had a comparatively large southerly component (Fig. 12b). That is, when the winds lack a southerly component, the EML cannot be entrained into the warm sector and no convergence line results, consistent with the analysis presented in the previous subsection.

5. Discussion

a. Limitations of QG diagnostics

In this study, we utilized QG theory to analyze the dynamics of prefrontal convergence lines. The prefrontal pressure trough in Fig. 1 is sharper than implied, for example, in Figs. 4, 8, and 9. This is not surprising given the relatively coarse horizontal resolution of the analysis fields as well as the smoothing, which was necessary to apply the QG framework. The smoothed forcing functions on such a coarse grid likewise reflect only the large-scale contributions. A possible interpretation of the results is thus that we captured the leading-order effect of the convergence-line formation. There may be additional effects that help collapse the boundary to smaller scales than what is diagnosed by QG theory. For instance, ageostrophic advection, which is neglected in QG theory, may yield more focused regions of forcing [e.g., within the context of semigeostrophic theory (Bosart and Lin 1984) or alternative balance (Davies-Jones 2009)]. Also, diabatic heating related to the evaporation of precipitation and solar heating varying across the convergence line (once clouds and precipitation have formed) could augment frontogenesis. Another possibility is that in some cases the collapse of the line does not rely on balanced dynamics altogether. Although this was not the case at the analysis time in Fig. 1, if linearly organized thunderstorms develop along the boundary, the leading edge of the outflow produced by the storms may become the primary, sharp convergence line.

The process of convergence-line formation proposed herein is reminiscent of the mechanisms described by Keyser and Carlson (1984) and Hanstrum et al. (1990a, b), who discovered regions of scalar frontogenesis focused along the edges of EML plumes.

b. Location of the convergence line relative to the EML

It is intriguing that the convergence line tends to be associated mainly with the western edge of the 850-hPa thermal plume (only at later stages when EML has deformed into a relatively thin filament does the convergence line move to the axis of the filament). To explore the role of deformation, Fig. 13 shows WRF output of the local dilatation axes at 850 hPa, as well as potential temperature.
As can be seen, the deformation magnitude (length of the line segments), the alignment of the deformation axes relative to the isentropes, and the horizontal \( \theta \) gradients are most favorable for scalar frontogenesis at the western periphery of the \( \theta \) plume. Reassuringly, this is consistent with the Q-vector configuration in Fig. 5. Further insight into the reconfiguration of the 850-hPa thermal ridge is provided by Fig. 14, which shows the evolution of two isentropes surrounding the thermal ridge for each of the simulated cases. Observe that the spacing between

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**Fig. 11.** Composite 850–700-hPa temperature lapse rates for cases (a) without and (b) with prefrontal convergence lines.

**Fig. 12.** Composite 850-hPa velocity vectors (\( \text{m s}^{-1} \), scale at the bottom of left panel) for cases (a) without and (b) with prefrontal convergence lines. The meridional component of the wind \( \nu \) is color shaded.
the isentropes as well as their orientation changes, especially at the western periphery of the thermal ridges, consistent with Figs. 5 and 13. This evolution is similar to the rearrangement of isentropes in highly idealized cyclone models (Doswell 1984; Schultz et al. 1998), in which isentropes closer to the vortex center (i.e., at the western periphery of the thermal ridge) are more strongly deformed than the isentropes at the vortex periphery (e.g., Fig. 12 in Schultz et al. 1998). As the thermal ridge becomes increasingly narrow, the maximum Q-vector convergence shifts to the thermal axis. This evolution, featuring Q-vector convergence in a narrowing warm tongue (Fig. 5), is quite similar to the occlusion process as described by Martin (1999b) and Schultz and Vaughan (2011). Since the EML is acted upon by the same wind field as the surrounding isentropes enclosing the warm sector, a possible interpretation of the convergence line is an “occlusion within the warm sector,” that is, a wrap-up of the thermal wave accompanying the EML [much as discussed by Schultz and Vaughan (2011)].

c. Relation to the warm season

An interesting question is why the prefrontal convergence lines are a warm-season phenomenon. The formation of the convergence lines hinges on the presence of an EML that needs to be drawn into the cyclone’s low-level circulation. During the cold season, the relatively small solar elevation angle does not support the development of a deep CBL over Iberia. The remaining EML source would primarily be the northern Sahara, but this source cannot easily be tapped, especially given that the cyclones’ tracks tend to be farther north in the winter than in summer. This is consistent with the composite analysis (section 4b), suggesting that in cases where the warm-sector flow does not extend sufficiently far into southwestern Europe, no EML is drawn into the eastern periphery of the cyclone.

6. Summary and concluding remarks

Using WRF simulations and Reanalysis-1 data, we addressed the dynamics of prefrontal warm-season convergence lines over western Europe. It was found that these lines are associated with an 850-hPa thermal ridge, which accompanies the base of an elevated mixed layer originating from the elevated terrain of the Iberian Peninsula and northern Africa. Quasigeostrophic theory qualitatively explains the existence of the convergence lines. As the isentropes at the periphery of the thermal ridge are deformed within the cyclone’s circulation, rotational and scalar QG frontogenesis occurs, which is accompanied by an ageostrophic circulation. The convergence line reflects the rising branch of this circulation. The QG forcing, which is strongly related to the horizontal deformation of the geostrophic flow, tends to be maximized at the western periphery of the 850-hPa thermal plume. Once the plume becomes sufficiently narrow after prolonged deformation, the forcing shifts toward the thermal ridge axis. At that stage, the configuration of the isentropes and the Q-vector forcing bear much resemblance to the occlusion process. This general pattern of evolution is supported by a 3-yr composite analysis. Here, we found that that the mid-level temperature lapse rates tend to be steeper and the 850-hPa flow more southerly, in warm sectors with a prefrontal convergence line than in those without. This analysis is consistent with the modeling results, sometimes, prefrontal rainbands occur over western and central Europe in the cold season, but those tend to be associated with split fronts (Pistotnik et al. 2011) accompanying vigorous extratropical cyclones.
highlighting the importance of EMLs in the formation of the convergence lines.

Figure 15 schematically summarizes the typical progression of the EML plume and its relation to the prefrontal convergence line. In Fig. 15a the cold front is still over the British Isles while a deep CBL is present over the Atlas Mountains and over the Iberian Peninsula (usually accompanied by a thermal low). About 12 h later (Fig. 15b), the CBL has been advected off of the Iberian Peninsula across France, where vector...
frontogenesis especially at the western edge of the EML leads to rising motion and a convergence line. Meanwhile, the cold front has made some eastward progress, and the EML is increasingly deformed while its distance to the cold front decreases (Fig. 15c). The thermal ridge associated with the EML becomes rather narrow, and the convergence line has shifted close to its axis. Finally, the cold front has almost caught up with the heavily deformed EML (Fig. 15d).

Further research is warranted to better understand the surface thermodynamic signatures of these lines, higher-order (than QG) effects that may contribute to the formation and intensification of the line, as well as the effect the circulation accompanying the line may have on the main cold front. Also, it would be interesting to explore EML-related prefrontal convergence lines over the United States.

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REFERENCES
Kaltenböck, R., 2004: The outbreak of severe storms along convergence lines northeast of the Alps. Case study of the


