The 9–10 November 2001 Algerian Flood

A Numerical Study

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On 9–10 November 2001, Algeria was struck with its most devastating flood since records were first kept in 1908 (Fig. 1). The flood was caused by an intense meso-β[1]-scale cyclone that struck the west side of the city of Algiers, producing accumulated rainfall up to ~285 mm and onshore winds of ~33 m s^{-1}, resulting in some 740 deaths. Fortunately, the overall cyclogenesis event was detected by the European Centre for Medium-Range Weather Forecast (ECMWF) model, enabling the Office National de la Meteorologie de Algier (ONMA) to issue a flood forecast as early as 5 November. Whereas the apparent predictability of the event suggests a controlling role by large-scale forcing, closer examination reveals that significant mesoscale development led to the actual weather pattern within the Algiers locale.

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FIG. 1. Devastation in Algiers, Algeria, on 11 Nov 2001 following flooding rains.

Figure 2 shows observed precipitation data provided to the authors by the ONMA for the event. The heaviest rainfall was focused along the north-facing slopes of the Atlas Mountains, west of Algiers, where

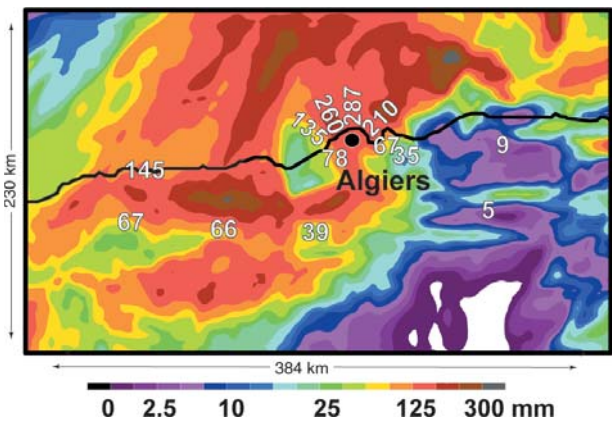

FIG. 2. Measured and simulated total accumulated storm rainfall along North African coast. Observed rainfall is indicated by white numerals (in mm) while UW-NMS rainfall on fine mesh grid after 48 h of integration is indicated by color shading (also in mm).

[1] Meso-α (200–2000 km), -β (20–200 km) , and -γ (2–20 km) scales as defined by Orlanski (1975).

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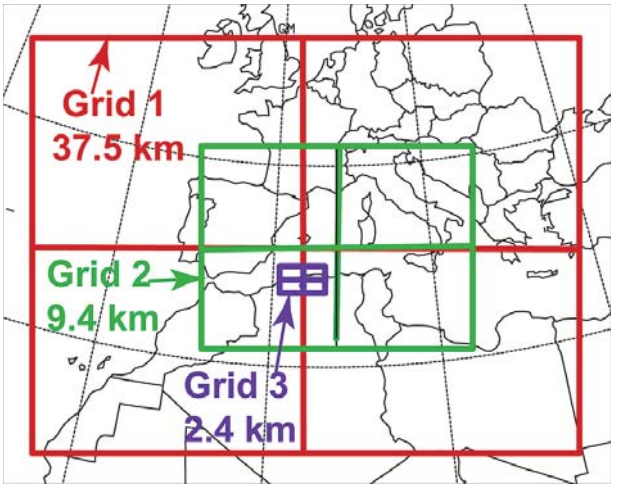


Fig. 3. Model grid setup showing position of three two-way interactive nests and their grid spacing. Grid design is cast on rotated spherical grid with origin at center of grid 1.

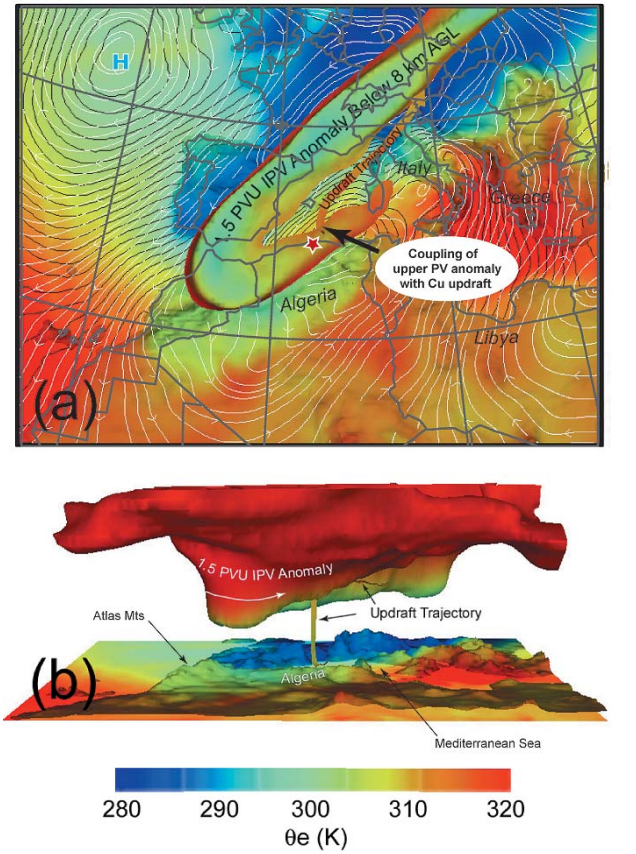


Fig. 4. Simulated surface of 1.5 potential vorticity unit (PVU; 1 PVU = 1.0 x 10^{-6} m^2 s^{-1} K kg^{-1}) IPV anomaly at 0800 UTC 10 Nov viewed from above in (a) and from south in (b). In (a): (1) position of city of Algiers is denoted with a red star; (2) IPV surface (color shaded according to its θ_e value) is cut off above 8 km MSL to reveal geometry of injected filament; (3) surface topography is shown as separate surface, shaded with surface value of θ_e; (4) wind streamlines at 1 km MSL elevation are shown in white, while MSL pressure is contoured in black (only over water) at 1-hPa interval; and (5) single trajectory characteristic of upward mass flux within convective plume is shown, shaded by θ_e values along the trajectory. When viewed from south in (b), the trajectory is seen flowing first upward and then northward after the equilibrium level is reached. Note how trajectory is initially colored equivalently to upper PV anomaly, consistent with coupling hypothesis.

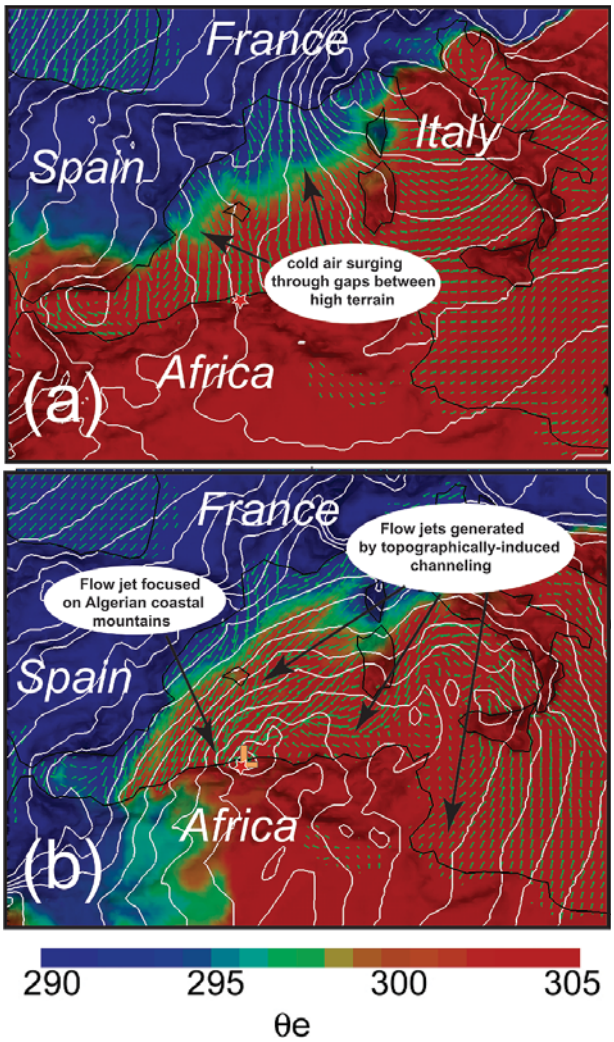
strong northerly winds moved onshore. To better understand the atmospheric evolution leading to the genesis of the storm, we performed a cloud-resolving model (CRM) simulation of the event using the University of Wisconsin Nonhydrostatic Modelling System (UW-NMS) (Tripoli 1992, 2006). The model was initiated with the 1° National Centers for Environmental Prediction (NCEP) Aviation Model (AVN) analysis at 1000 UTC 9 November and then integrated for a period of 60 h. Three two-way nested grids were employed as illustrated in Fig. 3.

RESULTS OF MODEL SIMULATION. The UW-NMS simulation was successful in reproducing the event's reported rainfall and winds. The simulated rainfall maximum around the city of Algiers was only slightly less than observations indicate (Fig. 2), and the distribution from the west to the east side of the city was captured well. Some degree of difference was not surprising since the principle sources for the heaviest precipitation included convective storms not anchored to topography. Such features are less predictable than topographically anchored orographic storms. The simulated winds were predicted well, with the UW-NMS's peak surface wind gust of 33 m s^{-1} near Algiers exactly matching peak gusts reported by ONMA.

The cyclogenesis process and its predictability had much in common with the 1993 "Storm of the Century" (SOC) event, which developed 12 March 1993 off the Texas coast, as discussed by Bosart et al. (1996). They showed the SOC to have been a predictable coupling between an upper-level isentropic potential vorticity (IPV) anomaly[2] to the surface-facilitated

[2] IPV refers to isentropic potential vorticity anomalies as discussed by Hoskins et al. (1985), which are analogous to potential vorticity (PV) anomalies relative to thickness-weighted averages along potential temperature layer (Ertel 1942).

Fig. 5. View of simulated low-level flow vectors on grid 2 (medium mesh) at 10 m MSL at (a) 0800 UTC 9 Nov and (b) at 1000 UTC 10 Nov. MSL pressure isobars at 2-hPa interval are also shown. City of Algiers is denoted with a red star. Topography surface is shown shaded in θ_e (K). Channeling around topography and transport of warm θ_e is evident in lower panel.



surface heating from the warm waters over the Gulf of Mexico. Similar to their analysis, we also employ "PV thinking"[3] to examine the synoptic forcing leading up to the Algeria flood.

In the days prior to the storm, our simulations reveal an exceptionally strong southwestward amplifying upper-tropospheric pressure trough. This produced an unusually deep tropopause fold, forcing stratospheric IPV values down to 700-hPa (Fig. 4). This scenario was nearly equivalent to the situation that had occurred in the western Gulf of Mexico leading to the SOC, as described by Bosart et al. (1996). Our analysis also suggests that the coupling of the surface air stream to the IPV filament aloft resulted in surface storm cyclogenesis over the Mediterranean.

Figure 5a depicts surface flow evolution as the system moved south off the coasts of Spain and France on the morning of 9 November, one day prior to storm genesis. Of interest is the extended influence of the upstream mountain ranges in channeling cold air movement. As cold surges reached the southern Mediterranean shores, an easterly low-level jet formed north of Tunisia and Algeria (Fig. 5b). We initially interpreted this feature as a barrier jet associated with the steep rise of the Atlas Mountains. However, subsequent impact experiments (described later) indicate that the jet formed independent of the mountains. The jet flowed westward, feeding into the developing upward ageostrophic circulation aloft, forming in advance of the approaching upper level IPV anomaly that was rounding the base of the trough from the southwest. As evident in Fig. 5b, over the 24-h period prior to cyclogenesis, Wind Induced Surface Heat Exchange (WISHE)[4] along the easterly low-level jet axis and over the warm waters north of Tunisia and Libya increased the boundary layer equivalent potential temperature (θ_e) values to match potential temperature (θ) values within the IPV anomaly aloft.

For coupling to occur between the warm surface θ_e and the IPV anomaly in the vicinity of Algiers, deep moist convection had to develop under the IPV filament. As evident in Figs. 6 and 7, simulated deep convection was initiated offshore on the morning of 10 November, with enhanced surface convergence forming beneath

[3] "PV thinking" denotes use of conservation and invertibility of PV to understand the dynamics of balanced flow.

[4] Wind-induced surface heat exchange (WISHE), as defined by Yano and Emanuel (1991), is based on positive interaction between a boundary layer circulation and heated surface. This mechanism is distinct from a Conditional Instability of the Second Kind (CISK) mechanism, the theory originally defined by Charney and Elliasen (1964) in which boundary layer circulations amplify through their interaction with convection in a developing warm core vortex. In contrast, WISHE theory requires surface heat fluxes to act as the principal mechanism for maintaining an energy source within an amplifying warm-core perturbation, with convection used only to distribute surface heat aloft.

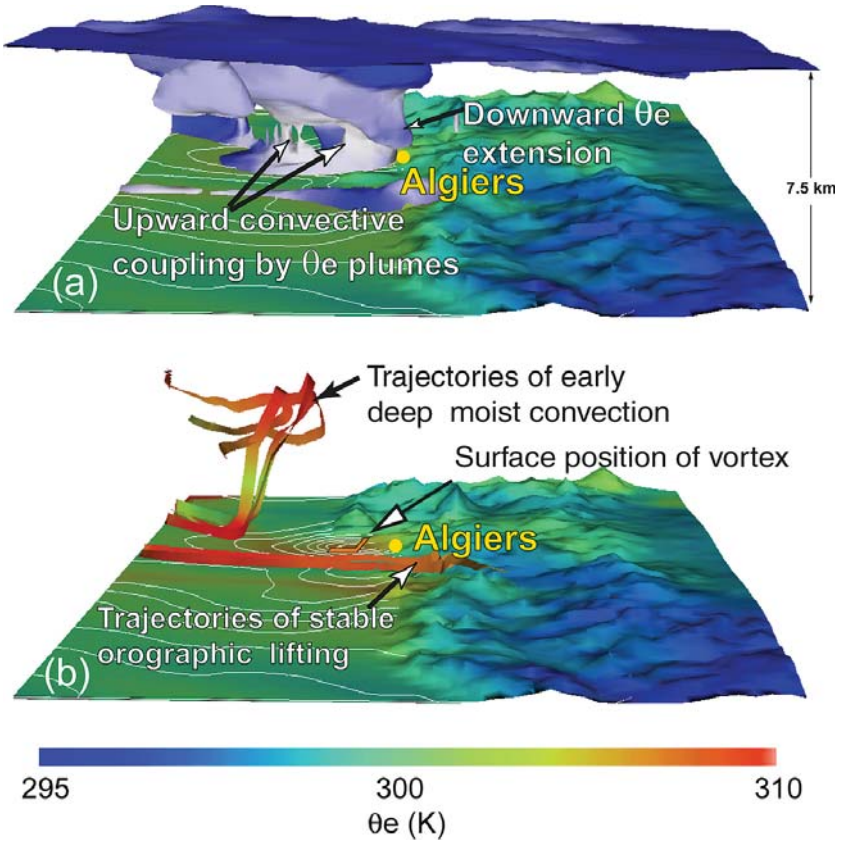


Fig. 6. View from south of simulated Algiers cyclone at 1000 UTC 10 Nov on fine-mesh grid 3. White contours are isobars of MSL pressure at 1-hPa interval. Trajectories in (b) are terminated at 1000 UTC. Trajectories and surface topography are color shaded according to θ_e, where the 306 K θ_e surface is shown in (a) and for which white shading indicates log of total water condensate.

the left exit region of the upper-level jet streak that was approaching from the southwest. The coupling of the surface to the warm IPV filament aloft is illustrated in Fig. 6a. Examination of the low-level jet evident at this time (Fig. 7) reveals that cooler postfrontal θ_e air sweeping off the African continent from the south converged into the warm low-level jet north of the African coast, producing the structure of occluding fronts with a tongue of secluded warm air sandwiched in between. The confluent secluded air subsequently rotated northward (Fig. 7b), wrapping around the developing low pressure center. The occlusion, in effect a bent-back warm front (Shapiro and Keyser 1990), exhibited strong horizontal shear featuring easterlies on the north side and southerlies in the postfrontal air to the south southwesterlies to the south.

It is notable that heavily precipitating airstreams forced into the Atlas Mountains southwest of the upper-level ageostrophic divergence (Fig. 6) did not result in simulated deep convective storms, simply because the orographic flows were not lifted high enough to achieve their vertical levels of free convection (LFCs). The contrast between the orographic flow trajectories and those in the low north of Algiers reveals the mechanism behind the storm's genesis. Over the water north of Algiers, WISHE acted to raise the surface humidity and temperature, effectively lowering the LFC. This led to the outbreak of deep convection over the water that formed a sudden direct coupling of the surface to the IPV anomaly aloft over the water.

Soon after the release of deep convection, the simulated surface vorticity began to strengthen dramatically, accelerating to levels in excess of 500×10^{-5} s^{-1}. This was aided by convergence of angular momentum as flow accelerated into the localized regions of convective lifting. In effect, there was "balling up" of the aforementioned vortex sheet along the bent-back warm front, a process described by Jascourt (1997). A surface low formed in the balled-up seclusion by 0300 UTC and subsequently deepened 10 hPa to 989 hPa by 0830 UTC. Northerly near-surface winds exceeding 37 m s^{-1} were simulated over the water on the western flank of the vortex. The resultant shallow warm core meso-β-scale vortex remained stationary along the shoreline on the northwest side of the city of Algiers for over 9 h. During this period, simulated surface winds peaked at 33 m s^{-1} within Algiers, while heavy rains, augmented by orographic effects, took place.

To better understand the roles of topography, land surface fluxes, and WISHE in forming and maintaining the meso-β-scale low, we performed three numerical experiments comprising a set of sensitivity tests, with results shown in Fig. 8 at the time of cyclogenesis on the morning of 10 November, some 34 h after the model run was initiated. The

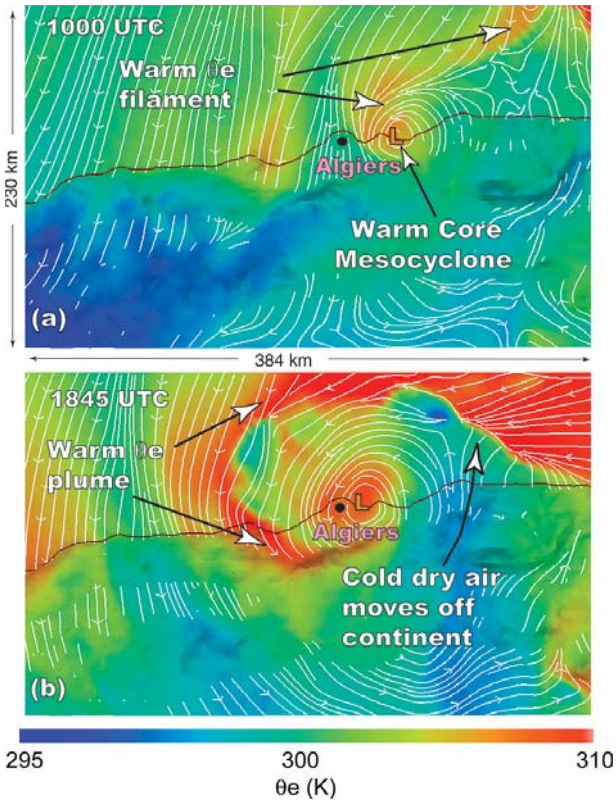


Fig. 7. Plan view of near-surface streamlines and topography surface on fine-mesh grid 3, shaded with θ_e at (a) 1000 and (b) 1845 UTC.

control experiment was designed identically to the original three-grid-nest run, but with only the outer grid included (Fig. 8a). (For computational economy, only the outer nest was used for these experiments.) For the first test (impact experiment 1), all African topography was removed while retaining the solid land surface (Fig. 8b). For the second test (impact experiment 2), African topography was retained but with the assumption that Africa was a water world (Fig. 8c). For the third test (impact experiment 3), surface heat and moisture fluxes were eliminated over the entire simulated surface (Fig. 8d).

Surprisingly, the results indicated only minor changes in the storm's genesis and precipitation pattern when African topography was removed (impact experiment 1). The most notable effect was a shift of the meso-β-scale low center westward, but by less than 50 km relative to the control experiment. In fact, the frontal system developed a more classic arc in the absence of mountains with a more coherent cold surge over Africa. Converting Africa to a water surface (impact experiment 2) also had only minor effects on the storm's genesis. In this case, the low developed eastward of the control position by about 50 km. The cold front was also slightly less coherent and intense, but still underwent a similar occlusion scenario. However, turning off surface heat and moisture fluxes (impact experiment 3) had a profound effect. The removal of WISHE eliminated the development of the meso-β-scale low altogether, leaving only the meso-α low associated with synoptic-scale positive vorticity advection aloft. There, the upper-level disturbance could couple to the surface on the basis of dry warm-sector θ values alone.

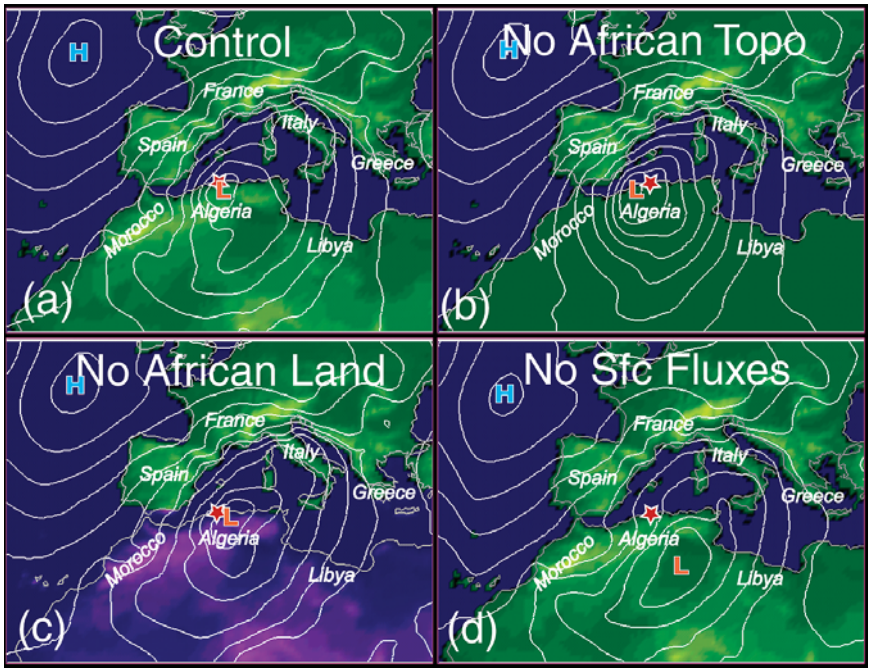


Fig. 8. Simulation of surface pressure field at 1200 UTC 10 Nov for control and three sensitivity experiments employing only the outer grid. City of Algiers is denoted by a red star. Topography is shaded, with higher elevations indicated by more yellow and purple hues. Note in (c), there is no African land surface, but topography variation is assumed in water surface. Surface pressure is contoured at 4-hPa interval with low- and high-pressure centers labeled.

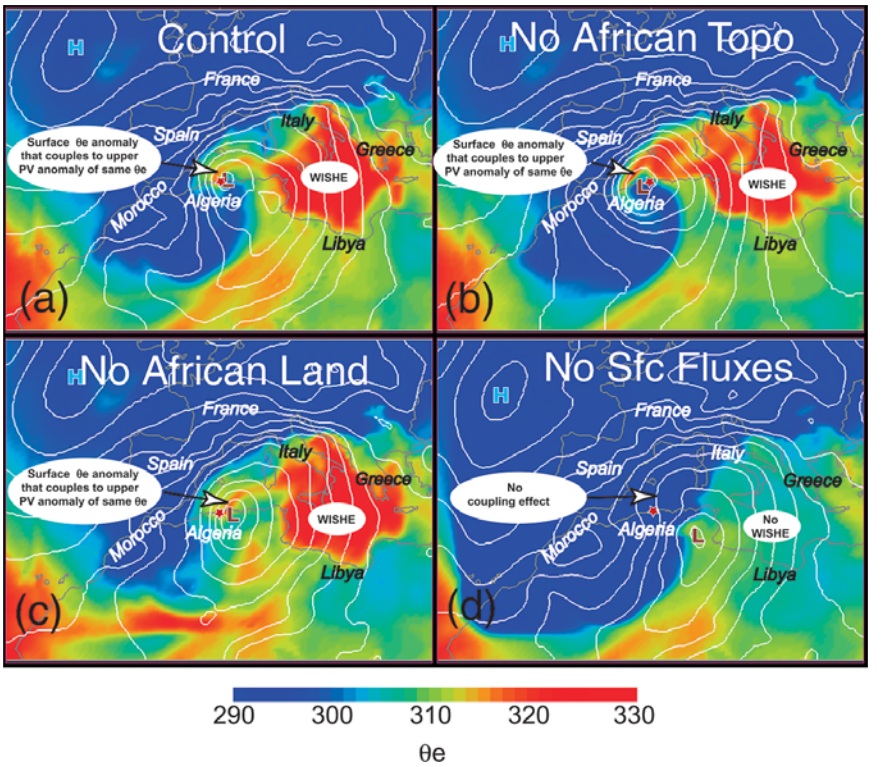


Fig. 9. Same as Fig. 8 except at 1800 UTC with shading performed on surface θ_e. Regions of strongest WISHE are depicted over the Mediterranean, north of Libya. Warm tongue of moist θ_e is seen wrapping around the north side of low in panels (a)–(c) in association with warm occlusion-like feature, a feature which does not exist in (d) because surface heat and moisture fluxes are suppressed, eliminating WISHE influence. Post-occlusion western extension of the warm θ_e tongue results in deep convection coupled to tropospheric injection of IPV aloft—also depicted in Figs. 4 and 6. Region of anomalous θ_e near boundaries, especially evident in (d), is the sponge region within which the UW-NMS model output is merged with NCEP-GFS output. This region appears because the lack of surface fluxes causes the UW-NMS solution to significantly depart from the NCEP-GFS solution.

Figure 9 depicts surface θ_e for all four experiments at 1800 UTC 10 November, elucidating underlying differences in the sensitivity tests. In the control and first two impact experiments, a tongue of high θ_e air is depicted, circulating around the northwest side of the meso-β low. In the absence of WISHE, Fig. 9d demonstrates no such tongue of warm θ_e air and no meso-β-scale low genesis north of Algeria.

The role of the warm seclusion in facilitating surface cyclogenesis through moist dynamic coupling is similar to the mechanisms leading to polar lows discussed extensively by the Shapiro and Fedor contribution to the Twitchell et al. (1989) compendium. In the storm cases they considered, the warm ice-free waters of the Arctic build a strong seclusion area that couples to the upper-level wave, leading to rapid cyclogenesis of a warm-core polar low. A three-dimensional visualization analysis of the Algerian storm reveals it to feature a shallow warm core structure within an otherwise much larger cold core structure associated with the upper-level wave, consistent with the polar low structures discussed by Shapiro and Fedor.

CONCLUSIONS. This study has found that the Algerian flood of 9–10 November 2001 resulted from convective coupling of an anomalously strong upper-level tropopause fold (and an associated strong IPV anomaly) with surface air warmed by WISHE over the warm southern waters of the Mediterranean Sea. This differs from classic Eady wave development not only because of the role of WISHE, but also because of the role of moist coupling of the surface with the upper-level wave via deep moist convection. This "conditional coupling" can lead to a more catastrophic, smaller-scale, and unusually rapid and intense cyclogenesis than one generally expects from more continuous and neutral dry vertical coupling. When occurring in the presence of a particularly strong upper-level IPV anomaly, the results can be devastating.

It has been shown that the dynamic mechanisms leading to the genesis of the Algerian flood were similar to those suggested for the SOC that occurred over the Gulf of Mexico about a decade earlier—which also had extreme consequences. Also, it has been shown that the storm's genesis and resulting low-level warm-core structure were similar to the genesis and structure of polar lows common to the North Atlantic ice shelf. Fortunately, as is often the case for Mediterranean storms, these local flow systems are predictable given a model's ability to finely resolve

topographical, sea-surface, land-use, and coastline features and skillfully simulate their interaction with the broader scale flow.

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FOR FURTHER READING

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