A Model Investigation of the Dynamics of a Coastally Trapped Disturbance

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ABSTRACT

Numerical simulations of the 15–17 May 1985 coastally trapped disturbance (CTD) event along the west coast of North America are compared with the schematic model of CTD evolution developed by Skamarock, Rotonno, and Klemp (SRK), which was based upon more idealized simulations. It is shown that the general evolution of the simulated May 1985 CTD is consistent with the SRK schematic model. It is further shown that secondary effects not contained in the SRK simulations, such as diurnal radiation variations and mesoscale topographic variations, can account for the variable CTD initiation and propagation observed both in nature and in the present numerical simulations. Diurnal radiation variations, coupled with differential heating of land and ocean, appear to play an important role in setting up the alongshore temperature gradient necessary for CTD formation and evolution. The modeled CTD is found to change dynamical characteristics from an initial Kelvin wave/bore similar to that discussed by Ralph, Nieman, Persson, Bane, Cancillo, and Wilczak to a gravity current, and this change is consistent and coincident with a sharp change in translation speed of the disturbance.

1. Introduction

Coastally trapped disturbances (hereafter CTDs) are mesoscale systems that are laterally confined against a coastal mountain barrier by the Coriolis effect and vertically confined by stable stratification. CTDs propagate along the coastal mountain barrier such that the barrier is on the right (left) in the Northern (Southern) Hemisphere. Propagation is generally energetic but relatively short-lived (typically 2 to 3 days on the North American west coast). Typical length scales are 1000 km alongshore, 100 km across-shore, and 0.5 km in the vertical (Reason 1994). CTDs along the west coast of North America are generally coastal ridges of higher pressure in the marine layer and result in a wind reversal from the more typical northerly to southerly flow (e.g., Dorman 1985, 1987; Mass and Albright 1987; Reason and Dunkley 1993).

There has been considerable dispute in the literature on the correct dynamical interpretation of CTDs. They have been interpreted as freely propagating Kelvin waves of various types (Dorman 1985, 1988), as trapped gravity currents (Dorman 1987; Mass and Albright 1987), and as the forced mesoscale response to purely synoptic-scale evolution (Mass and Albright 1988; Mass and Bond 1996; Mass 1995). Using an analytical shallow-water approach, Reason and Steyn (1992) address the relative importance of boundary layer processes (i.e., Kelvin wave and gravity current dynamics) and synoptic-scale processes. A recent analysis of more detailed observations from the 10–11 June 1994 event (Ralph et al. 1998; Ralph et al. 2000) identified the CTD response as a mixed Kelvin wave/bore occurring in the inversion layer above the marine boundary layer (MBL). A bore is in some ways similar to a gravity current. One difference is that the zone of strong thermal gradient in a bore occurs aloft and does not extend to the surface. Another difference is that while the translation speed of a gravity current must be less than or equal to the wind component in the direction of propagation within the gravity current, the phase speed of a bore can be greater than the wind speed in the direction of propagation (Ralph et al. 2000).

In their synoptic climatology, Mass and Bond (1996) found that offshore (easterly) flow at 850 hPa was a robust feature accompanying the initiation stages of CTDs. In order to examine the importance of offshore flow on CTD initiation and propagation, Skamarock et al. (1999, hereafter SRK) utilized a nonhydrostatic numerical model in 2D and 3D simulations with idealized
topography and background synoptic conditions. The model used was inviscid and purely dynamical—it did not include surface fluxes, diurnal radiation effects, or offshore advection of heated air. They examined the formation and propagation of a CTD resulting from the gradual introduction and subsequent removal of a tapered zone of offshore flow over the middle of their model domain under different marine boundary layer conditions and stratifications. They found that the offshore–downslope flow can lead to adiabatic warming and evacuation of the MBL near the coast causing formation of a surface mesotrough. This results in a displacement offshore of the climatological northerly jet and in onshore flow to the south of the mesotrough feature. The onshore flow to the south results in convergence of marine air and an increase in the MBL depth causing the formation of a mesoridge along the coast. The mesoridge is accompanied by southerly flow and propagates to the north as Kelvin wave–like CTD, which decays exponentially offshore. This model of CTD evolution is depicted schematically in SRK’s Fig. 17 (reproduced here as Fig. 1). Their model was shown to be in qualitative agreement with observations and simulations of the 10 June 1994 event (Ralph et al. 1998; Thompson et al. 1997) including the presence of southerlies above the MBL, elongation of the mesotrough, and intensification of the northerly jet offshore.

SRK also found that with stratification above the MBL, a topographically trapped Rossby wave could form that propagates northward above the MBL modifying the CTD response. The topographic Rossby wave resulted in pressure changes outside the MBL that could be mistakenly interpreted as synoptic evolution continuously forcing the CTD. In summary, SRK provide a convincing model for the general evolution of CTDs and for their interpretation as essentially freely propagating Kelvin waves.

In this study, we reconsider the CTD event of 15–17 May 1985 along the west coast of North America. This event, originally documented by Mass and Albright (1987), represents an example of a strong mesoscale-trapped event with abrupt transitions in many basic meteorological parameters. Its propagation along the Oregon–Washington coasts has previously been inter-
Interpreted as a trapped gravity current (Mass and Albright 1987; Jackson et al. 1999). According to the Bond et al. (1996) 11-yr CTD climatology, the southerly flow during this event was of the longest duration of all events in southern California, although the southerly flow speed maximum in southern California was average to below average. Along the Oregon coast, the southerly flow duration for this event was average, although the maximum strength of the southerlies there was the second strongest on record.

Initiation of the May 1985 CTD occurred with the following synoptic pattern developments (Guan et al. 1998). A northeastward tracking high pressure system offshore of Oregon moved inland into southwestern British Columbia while a continental low pressure system tracked to the southwest toward southern California [see Guan et al. (1998) Figs. 4 and 6]. The resulting pressure gradient set up northeasterly (offshore) geostrophic flow along most of the western U.S. coastline. Westward movement of the continental low combined with lee troughing led to the formation of a coastal trough midway along the California coast. Near-surface flow adjustment to this synoptic evolution included the offshore migration away from the coast of the climatological northerly flow, and onshore flow convergent onto the coastal mountains to the south of the low. A decrease in the MBL depth (referred to as MBL suppression) by the offshore flow to the north [also seen for other cases in Thompson et al. (1997) and SRK] and the raised MBL to the south due to convergence on the barrier produced a “step” in the MBL and a near-shore coastal ridge, which began propagating northward as the CTD. The across-shore scale, offshore decay and horizontal trapping of the May 1985 CTD are indicated by the mesoscale pressure analyses presented in Mass and Albright (1987, their Fig. 9) and reproduced in Guan et al. (1998) and by the RAMS simulated sea level pressure fields in Guan et al. (1998, their Fig. 8). For a structural snapshot of the mature CTD, refer to Fig. 2, which depicts the modeled sea level pressure and the 216-m elevation, wind, potential temperature, and mixing ratio, as well as an IR satellite image. At 0000 UTC 17 May, the disturbance is just south of the Columbia River; it is marked by a narrow coastal ridge with southerly flow in Fig. 2a. A strong thermal gradient is evident at the leading edge of the disturbance (Fig. 2b) with its offshore scale and decay apparent from Fig. 2a. The close association between the simulated moisture field and the observed satellite cloud pattern can be seen by comparing Figs. 2c and 2d. The trapped nature of the disturbance is evident from the offshore decay. In most respects, the observations and simulation of the May 1985 CTD are consistent with the model of evolution proposed by SRK. In the SRK simulations however the CTD was initiated by a stationary pulse of offshore flow introduced and then removed in the center of their domain. Observations and Regional Atmospheric Modeling System (RAMS) simulations from the May 1985 event indicate that the zone of offshore flow propagated northward ahead of the disturbance and thus could provide continual forcing.

Validation and analysis of simulations of this particular event using the Colorado State University RAMS has been presented in previous papers (Guan et al. 1998; Jackson et al. 1999), and details of the model and its configuration can be found there. The RAMS simulations are not as idealized as the SRK simulations—they are made using a model that includes realistic topography, surface fluxes, and diurnal radiation variations. Additionally, the lateral boundaries of the RAMS simulations on the coarsest grid are continuously nudged toward NCEP analyses, so that synoptic-scale variations are included in the mesoscale model. The disadvantage of using a model like RAMS, compared with a more simplified modeling approach such as that used by SRK is that interpretation of the underlying dynamics is more complicated—it is more difficult to draw conclusions from the model results because of the potentially confounding effects of many simultaneous processes. The major advantage, however, is that RAMS is able to represent processes not contained in simpler models, which may be important to the evolution of CTDs.

In the present study, we discuss how the RAMS model results of the strong 15 May 1985 event compare with the general model of CTD evolution presented in SRK. We also show how factors not included in the relatively simple numerical models used in SRK, such as topographic variability and diurnal radiation variations with associated diabatic and advective effects, lead to changes in how CTDs are manifested, and are related to the unsteady propagation of these events observed in nature. It is further suggested that the 15 May 1985 event is best described as a mixed Kelvin wave/bore initially and then becomes more like a gravity current in its final stages, and that this change in dynamical character can explain the reduced propagation speed seen in RAMS simulations. This change in CTD dynamical behavior in the model is also consistent with surface observations from several studies of these phenomena (e.g., Mass and Albright 1987).

2. Methods

The RAMS modeling system is described fully in Pielke et al. (1992). The RAMS configuration used for the simulations discussed here are described more fully in Guan et al. (1998) but will be briefly summarized here. The model used was nonhydrostatic with horizontally variable initial conditions and time-dependent lateral boundary conditions from National Centers for Environmental Prediction (NCEP) analyses on the coarse grid. All simulations used the same NCEP analyses as initial conditions and to nudge the lateral boundaries on grid 1, so all contained the same synoptic forcing. Moisture was present in the model and could condense as cloud, but no ice or precipitation mechanisms
were present (precipitation was not observed during the event). Solar and terrestrial radiation were permitted on all grids (except the runs in which the radiation was turned off) and computed every 1200 s. An Arakawa type C grid stagger was used for the two nested grids—grid 1 at 100-km horizontal resolution, and grid 2 at 25-km resolution. A third nested grid, at 6.25-km horizontal resolution was experimented with but not found to lead to further insights at the scale addressed here. All grids had 32 vertical levels with a spacing of 75 m near the surface stretching by a factor of 1.2 for each successive level until a maximum separation of 1000 m was reached, resulting in a vertical domain of 16.5 km.
m above ground level for the other model variables. In order to determine whether 75-m vertical resolution was sufficient within the boundary layer, we tested the model with 40 levels starting at 25-m resolution for 36 h of simulation time, and found very similar evolution and MBL representation.

In order to reduce the confounding influence of topographic variability, and to enable a closer comparison with the SRK simulations, runs were made using a highly simplified representation of topography in which the Coastal Mountains and Cascades are combined into one range. Figure 3a shows the grid 2 uniform topography used, where western North America is represented as a 1500-m high plateau sloping uniformly to a coastline in three straight segments. Figure 3b shows the realistic grid 2 topography used in the other simulations.

In this paper we make use of six different simulations and a sensitivity simulation of the 15–17 May 1985 CTD event, which occurred along the west coast of North America (summarized in Table 1). In order to find the importance of realistic topographic variation not considered in SRK, the first three simulations utilize a uniform ramp topography (named with the prefix “U’’), while the last three utilize a realistic representation of topography (named with the prefix “R”). To find the importance of diurnal diabatic effects on CTD evolution, for each of the two topographies three simulations are made: 1) normal radiation cycle (suffix “rad”), 2) radiation offset by 12 h (suffix “rad + 12”), 3) no radiation (suffix “norad”). The simulation with realistic topography and a normal radiation cycle is the same as that presented in Jackson et al. (1999). That paper, together with Guan et al. (1998), provides an extensive validation of the ability of RAMS to represent the event. In this study, we analyze the output from the various simulations to further elucidate the fundamental dynamics present in this case.

In comparing the details of various simulations, we have found it very helpful to examine time sequences of the coast-parallel horizontal potential temperature gradient. The leading edge of a CTD is associated with a maximum horizontal temperature gradient (Fig. 2b). Further, the value of the maximum coast-parallel horizontal potential temperature gradient is related to the steepness and nonlinearity of the leading edge of the disturbance. In Figs. 4 and 5 we take sequences of the horizontal potential temperature gradient along the offshore slice defined by the line ABCD in Fig. 3 (parallel to the y axes of Figs. 4 and 5) at different times into the simulation (time increasing along the x axes of Figs. 4 and 5) at model level 4 (216 m above sea level, hereafter MSL) and contour these values. CTD are generally identified by the wind reversal (i.e., the change from northerly to southerly flow) but, as will be discussed below, the Urad flow (particularly above the surface) had a southerly component prior to the arrival of the CTD and hence the wind reversal cannot be used to determine the position of the CTD. We find that the surface wind reversal can be a good indicator of the CTD position, especially in the
south, for the realistic terrain simulations, and the simulations with radiation, but the alongshore potential temperature gradient is a more consistent indicator for all simulations. For this reason time sequence plots of alongshore potential temperature gradient (Figs. 4 and 5) and change in alongshore wind were constructed to gain further insight into the CTD propagation and evolution. The surface wind reversal position is indicated as a heavy dashed line on the potential temperature gradient plots (Figs. 4 and 5). It should be noted that the contours of maximum gradients presented in these plots represent the intersection of the MBL inversion (hereafter MBLI) with the $Z = 216$-m level. Thus, a near horizontal MBLI will appear as a weak gradient in potential temperature and alongshore wind, and a more vertical MBLI will be represented by stronger gradients. Using a hydrostatic argument the steeper MBLI will be associated with stronger gradients in surface pressure, assuming the steep MBLI exists over a nontrivial depth. Any MBL disturbances that exist entirely above or below the chosen level (216 m) will clearly not exist in these plots. The 216-m level was chosen because it captured the most detail of the MBLI evolution. The resulting figures summarize evolution, intensity, and propagation speed for an entire CTD simulation in one figure, facilitating comparison of differences between simulations. The CTD leading edge along the coast and its movement through time is marked by the locus of maximum contoured values. The slope of the locus of maximum contoured values corresponds to the translation speed—the more nearly vertical the faster the translation. The higher the contour value, the sharper the gradient at the leading edge, and presumably
the more nonlinear the disturbance. This analysis and interpretation of the CTD differs somewhat from other studies, where the focus has been on examining the wind reversal (which is observed) rather than the zone of strong thermal gradient associated with the disturbance, which cannot be analyzed from routine observations since the data do not exist. The wind reversal and zone of strong thermal gradient are generally related as can be seen in Figs. 4 and 5, however, as discussed later, it is possible for the southerly transition to occur ahead of this zone (Fig. 6a).

3. Uniform topography

Figure 4 shows the contour plots of coast-parallel potential temperature gradient time sequences for the three uniform terrain simulations. The locus of maximum coast-parallel potential temperature gradient indicates the time–space location of a “step” in the MBL (deeper to the south, shallower to the north) representative of the CTD location. There is a distinct change in translation speed in these figures, indicated by the change in slope of the locus of maximum coast-parallel potential temperature gradient. The early, more rapidly propagating part (i.e., steep slope prior to hour 48 in Fig. 4a) we call phase I, and the later, less rapidly propagating part we call phase II of the CTD.

Figure 4a is the reference uniform terrain simulation (Urad). By following the locus of maximum coast-parallel potential temperature gradient (i.e., the location of the MBL “step”), it is apparent that the CTD begins to initiate near simulation hour 12 at 200 km along the
section and propagates northward for about 8 h before a climatological strengthening of the northerly flow component due to daytime heating of the continent in the late afternoon backed the offshore flow ahead of the CTD into a more northerly direction and stalled the propagation (refer also to the strong northerly flow evident in Fig. 6). Overnight (after hour 30, 0600 UTC 16 May), the flow to the north of the CTD veered anticyclonically, returning to an offshore direction, strengthening the coastal lee troughing and resulting in more rapid northward CTD propagation as the CTD accelerated rapidly into the coastal trough. During the daytime and early evening hours (hours 18–28) when the CTD is stalled by the strengthened northerly flow, the CTD leading edge lags the coastal pressure minimum by as much as 300 km (see Jackson et al. 1999, their Fig. 5). Overnight, during the period of rapid propagation, the CTD “catches up” to the coastal pressure minimum. This behavior is consistent with the results of Thompson et al. (1997), who showed that the northward extent of coastal wind reversal (of a different case simulation) was limited by the position of the leading pressure trough. It is also consistent with SRK, who noted that a characteristic of all CTD is that the pressure and along-coast wind are generally in phase. In the Urad simulation, this phase relationship was disrupted by diurnal effects, and returned after the CTD accelerated ageostrophically into the coastal trough.

The average translation speed during this phase I of evolution, determined from the average slope of the locus of maximum coast-parallel potential temperature between simulation hours 30 and 40, is 20.1 m s$^{-1}$, and determined from the surface wind reversal is 18.0 m s$^{-1}$. These propagation speeds for the Urad simulation are faster than the 5–14 m s$^{-1}$ estimated from observed cloud motion and the wind reversal by Mass and Albright (1987). The Uniform barrier height is higher than the actual barrier height, especially farther north, which as discussed in Tory et al. (2001, hereafter TJR) would lead to faster propagation. Beginning at hour 48 at 1300 km along the transect (phase II), the coast-parallel potential temperature gradient increases and its slope (i.e., the disturbance translation speed) decreases to 9.0 m s$^{-1}$ (defining the CTD position as that of the maximum potential temperature gradient). The transition from phase I to phase II is discussed further below.

Figure 7 shows coast-parallel vertical cross sections of potential temperature and horizontal wind along the heavy line in Fig. 3a at 6-h intervals starting at simulation hour 24. Note the strong offshore flow accompanied by lowering of the MBL depth and warming above the MBL, which develops between hour 24 (Fig. 7a) and 30 (Fig. 7b) over the 500–1500-km region along the transect. This zone of strong offshore flow also exists prior to hour 24 (plots not shown) and acts to reverse the temperature and pressure gradients in a manner similar to that which SRK show schematically in Fig. 1. At this time, the CTD, as represented by the zone of strong horizontal potential temperature gradient, is lo-
Fig. 7. Coast-parallel vertical slices of potential temperature contours and horizontal wind as vectors for the uniform terrain simulation summarized in Fig. 4a. The vertical slice is along the heavy line ABCD in Fig. 3a. Up/downward pointing wind vectors indicate off/onshore flow perpendicular to the slice; right/left-pointing wind vectors indicate south/northerly flow parallel to the slice: (a) 0000 UTC 16 May 1985, (b) 0600 UTC 16 May 1985, (c) 1200 UTC 16 May 1985, (d) 1800 UTC 16 May 1985, (e) 0000 UTC 17 May 1985, (f) 0600 UTC 17 May 1985, (g) 1200 UTC 17 May 1985, (h) 1800 UTC 17 May 1985.

located well to the south with the leading edge between 200 and 300 km along the transect, and with onshore flow to the south of this location. Note that the nearly vertical isentropes (e.g., 288 K; Figs. 7a,b) marking the leading edge of the CTD are in the inversion layer above the MBL and that these isentropes do not descend to the surface, which is suggestive of a bore or mixed Kelvin wave/bore structure. The MBL does however increase in depth behind the CTD leading edge, differing somewhat from the 10–11 June 1994 event (Ralph et al. 2000) in which the CTD near the leading edge existed in the inversion above the MBL with the MBL depth itself increasing farther to the south.

The coastal surface pressure decreases in response to the offshore flow ahead of the CTD. In SRK, the offshore flow resulted in an evacuation of the MBL, which in combination with adiabatic warming acted to lower the coastal surface pressure. In our simulations, although the MBL depth is reduced in the offshore flow, a shallow MBL is maintained due to diabatic and frictional effects over the ocean. The MBL depth reduction and surface pressure decrease can be due both to en-
trainment of warm offshore flow with cyclonic potential vorticity (PV) into the marine layer (e.g., Reason and Jury 1990) and to a suppression of the MBL discussed more fully in TJR. In addition to adiabatic warming, the advection above the MBL of the radiationally heated continental boundary layer also contributes to lowering of coastal surface pressure in our simulations (although not in the SRK simulations). In order to assess the relative contributions of each mechanism to decreasing the surface pressure ahead of the CTD, we examined the Urad model atmosphere at 1000 km in the along-coast cross section at 18- and 30-h model simulation time. The atmosphere was divided into three layers: 0–700 m, 700–3300 m, and above 3300 m. In each layer the hydrostatic contribution to the change in surface pressure between these times was assessed. The changes in the lowest layer are due to a decrease in the MBL depth and account for a 3-hPa decrease in surface pressure. The warming in the second layer is due to the offshore advection of the well-mixed and heated planetary boundary layer (PBL) and account for a 4 hPa decrease in surface pressure. The changes in the upper layer account for a 3-hPa decrease in surface pressure and are attributed to synoptic-scale changes. Thus it is primarily the combination of a reduction in the MBL depth and warm advection above the MBL in the offshore flow that results in the coastal lee trough ahead of the CTD. This lee trough reverses the alongshore pressure gradient to the south resulting in ageostrophic northward acceleration of the CTD.

After hour 30, the CTD preceded by strong offshore flow and followed by southerlies with an onshore component and increased MBL depth, rapidly propagates to the north (phase I). During this period of rapid propa-
gation the disturbance deepens and the leading edge steepens (compare Figs. 7c and 7d), and the width of the disturbance, at the $Z = 216$-m level, increases from about 150 to 250 km (not shown). At approximately hour 48 (Fig. 7e), the leading edge of the CTD steepens even further and the gradient increases (as seen in Fig. 4a) with the nearly vertical isentropes extending to the surface, suggesting a gravity current structure. At the same time, the translation speed decreased—this is called phase II.

The transition to a gravity current–like structure begins with the steepening of the CTD leading edge in combination with the maintained MBL suppression ahead (Figs. 7d,e) and is completed when the strong thermal gradient extends to the surface (Figs. 7e,f). The extension of the thermal gradient to the surface is enhanced during the afternoon and early evening by the convergence (which also extends to the surface) between the strong daytime opposing northerly flow and the strong southerly trapped flow (Fig. 7e). In the CTD climatology of Bond et al. (1996), it was found that the leading edges of CTDs, as inferred from surface marine buoy observations, are sharper at more northern locations, suggesting wave steepening near the surface and a transition to gravity current characteristics. Also, Reason and Steyn (1992) suggest that observed abrupt transitions in CTD translation speed could be due to nonlinear wave steepening. SRK also report nonlinear steepening of the disturbance at the leading edge, which in their simulations occurs shortly after initiation in their experiments with a MBL present due to complete evacuation of the MBL by the offshore flow. The general evolution of this event, is very similar to that described by SRK and summarized schematically in Fig. 1.

SRK found a Rossby wave mode propagating along the Cascade Mountains above the MBL when they set the Brunt–Väisälä frequency ($N$) aloft to $10^{-2}$ s$^{-1}$. There was no Rossby wave mode in SRK with neutral stratification above the MBL. Stratification from the RAMS simulations in the layer between 1000 and 4000 m was of similar order, and a Rossby wave mode was present. TJR showed that the model Rossby wave mode was the dominant mechanism responsible for the propagation of a trough, confined to the coast, that led the CTD, and identified that this trough had a significant influence on the CTD propagation. This was particularly true during the phase I propagation when strong ageostrophic acceleration toward the leading trough minimum gave rise to rapid CTD propagation.

In summary, the CTD evolution for the uniform terrain simulation depicted in Figs. 4a and 7, seems to agree quite well with the more idealized simulations in SRK, which were summarized schematically in Fig. 1. Both results show a reduction in the MBL depth ahead of the CTD due to the strong offshore flow. However, in the SRK simulations the MBL is completely removed by the offshore flow since there are no surface fluxes present to maintain it, while this layer is only suppressed in the present simulation. In addition, the SRK simulation does not allow for offshore advection of air heated during the daytime to supplement the adiabatic warming due to offshore flow in the coastal zone, which is deemed important in the present simulation. Since these two limitations of SRK offset each other, the overall evolution remains similar to the more realistic simulation presented here. Interpretation of the present simulations, which will be supported by an analysis of propagation speeds presented in the discussion section, indicates that the event has characteristics of a bore or mixed Kelvin wave/bore initially (phase I) and that a transition to a gravity current (phase II) occurs during the event evolution.

4. Diurnal diabatic effects

In order to assess the importance of diurnal diabatic effects, which were not included in the SRK model, three additional simulations were made: one with the radiation cycle offset by 12 h (Urad + 12 - Fig. 4b), the second with radiation turned off altogether (Unorad, Fig. 4c), and the third with normal radiation but the fluxes of heat and moisture between the land and atmosphere turned off (Unoheat). In Figs. 4a and 4b the shading indicates 8 h of peak solar radiation heating centered upon local noon. By comparing Figs. 4a with 4b one can see quite significant differences in evolution caused by differences in the timing of the solar heating cycle. Specifically, the coast-parallel potential temperature gradient at the CTD leading edge seems to weaken during times of peak heating (e.g., Fig. 4b between hours 28–36 and 52–60), and the translation speed also seems to decrease during these times. The decreased daytime translational speed in the Urad + 12 simulation is less pronounced than in Urad (Figs. 4b,c) due to the later application of the heating cycle. By the time the heating was applied (between 28 and 36 h) the lee troughing that led the wave was considerably more developed than in the Urad simulation, which led to a further offshore migration of the opposing northerly flow, and consequently weaker resistance to propagation during the day. The reduced stalling is also a result of the increased strength of the trapped southerlies at this time, compared with the Urad simulation. In the Urad simulation (Fig. 4a) the potential temperature gradient generally decreases or is approximately constant during peak heating times, except after the middle of the heating cycle between hours 45 and 48, when there is a slight increase. Radiation is important in developing the coast-parallel potential temperature gradient—in Fig. 4c in which the radiation is turned off—the CTD leading edge has a much smaller potential temperature gradient and the disturbance translation speed is generally less. Solar radiation acts to increase the coast-perpendicular thermal gradient near the surface by differentially heating the continental surface compared with the marine surface. In the presence of offshore flow in the north
this results in a northward-pointing coast-parallel thermal gradient since the warmed continental air is advected over the coastal zone. In other words, differential offshore warm air advection leads to rotation of the alongshore potential temperature gradient from the normal southward-pointing direction (i.e., warm air in the south) to the northward-pointing (i.e., warm air in the north) direction. The importance of diurnal diabatic effects on the evolution and intensity of CTDs is highlighted by the dramatic differences seen between Figs. 4c and Figs. 4a,b—these diabatic effects are not included in SRK or in simpler numerical or analytic models of CTD.

Turning off the radiation in the Unorad simulation has the combined effect of shutting down the diurnal heating of the atmosphere by the surface in response to daytime solar radiation thus decreasing warm offshore advection to the north, and of shutting down cloud-top radiative cooling in the cloudy atmosphere within the CTD. Both effects contribute to a weakening of the alongshore potential temperature gradient. Figure 2c shows mixing ratio at 216-m MSL, which corresponds quite closely with the satellite imagery in Fig. 2d, lending confidence that the moisture pattern was reasonably well simulated in the model so that IR radiative cooling from the cloud tops should also be contained in the simulations with radiation. In order to see whether long-wave radiative cooling in the cloudy atmosphere might be an important contributor, we have made an additional simulation (Unoheat) in which we left the radiation turned on, but turned the soil model off (so that there could be no heat or moisture fluxes into the atmosphere from the land). In this simulation, in which longwave cooling from the cloud tops could exist but diurnal continental heating of the atmosphere could not, the MBL within the CTD was about 2 K cooler, but the CTD evolution and strength of the potential temperature gradient was essentially the same as the Unorad simulation. We conclude from this that longwave radiative cooling from the cloud tops is of secondary importance in forcing the CTD compared with offshore warm advection of continentally heated air.

The change in translation speed between phase I and phase II, coincident with the nonlinear wave steepening and transition from a bore to a gravity current, which is denoted by the change in slope of the locus of maximum coast-parallel potential temperature gradient, depends on the timing of the diurnal solar cycle. Even with the same synoptic forcing, the transition that occurs at the onset of nighttime near hour 48 in simulation Urad (note the change in slope of the locus of maximum horizontal potential temperature gradient at this time in Fig. 4a) is delayed by approximately 7–14 h in the Urad + 12 simulation (the change in slope now appears to occur at hours 55–62 in Fig. 4b). In the Unorad simulation (Fig. 4c) nonlinear wave steepening and a transition from a bore to a gravity current do not appear to occur and the disturbance propagates more steadily. This occurs for several reasons. In the Unorad simulation the coastal trough is not as deep since there is no contribution to the troughing from the advection of radionally heated continental air by the offshore flow. Therefore the ageostrophic acceleration of the flow into the trough is less and the coast-parallel thermal gradient is less so that northward propagation is slower and nonlinear steepening does not occur. Also, in the Unorad simulation, the MBLI strength is much weaker since there is no offshore advection of radionally heated air above the MBLI to intensify it. Consequently the Unorad system behaves much less like a two-layer system and more like a weakly continuously stratified system so that shallow-water wave dynamics (such as nonlinear wave steepening) are less important.

The impact of the diurnal solar radiation cycle on CTD evolution is multifaceted. During the day, heating over the land lowers pressure there and enhances the opposing northerly coastal jet decreasing the CTD translation speed. This diurnal variation of the northerly coastal jet along the California coast has been previously documented in Beardsley et al. (1987) and discussed further in Ralph et al. (2000). Offshore advection of radionally heated air ahead of the CTD warms the air to the north and is largely responsible for the coast-parallel temperature gradient and hence the coast-parallel pressure gradient, which forces the CTD. In SRK these gradients were created somewhat artificially without offshore warm advection, by complete evacuation of the MBL by the offshore flow and by adiabatic warming of the air as it flows down the coastal mountains. In the present study, in addition to adiabatic warming and a decrease in the MBL depth, it is the offshore advection of continentally heated air that further lowers the coastal surface pressure and promotes the transition from a largely elevated Kelvin wave/bore disturbance to a surface-based gravity current during phase II. These two impacts of the diurnal cycle tend to counteract each other, with enhancement of the northerly jet slowing and weakening CTDs while advection of warm air offshore enhances CTDs. The offshore advection of warm air and adiabatic warming together were found to account for 40% (4 hPa of the total 10 hPa) of the decrease in coastal surface pressure that forced the southerly flow in the simulation. The observed northerly jet in the layer below 1200 m was examined by Ralph et al. (2000), who found the average jet of 1.3 ± 0.3 m s⁻¹ during the night. We calculate a daily average northerly jet in this layer to be 2.7 m s⁻¹ reaching an hourly average maximum of 5.5 m s⁻¹ at 0200 UTC (18 LST). Given the propagation speeds estimated from the Urad simulation, of 20.1 and 9 m s⁻¹ in phase I and II, the daily average retarding effect of the northerly jet on CTD translation is 12%–23%, reaching a maximum of 21%–38% at 0200 UTC. The two impacts are however not entirely in phase, with enhancement of the northerly jet occurring from midnight to late evening (Ralph et al. 2000, their Fig.
7) while offshore advection of air heated during the day occurs during the night as well. This favors northward CTD propagation between late evening and late morning.

The transition between phase I and phase II and the presence of gravity current–like propagation occurs during times of diminished solar radiation—the nonshaded times in Figs. 4 and 5—whereas propagation as a bore during phase I can occur during times of peak solar radiation intensity. This also lends support to the idea that phase I propagation is like a bore or mixed Kelvin wave/bore with the zone of strong thermal gradient located aloft where it cannot be as easily affected by solar radiation. The elevated nature of the phase I propagation can be seen by referring to Figs. 7a–c in which the strongest horizontal potential temperature gradient exists in the MBLI and does not extend to the surface. During phase II propagation as a gravity current, the horizontal change in pressure across the gravity current leading edge near the surface can be diminished by solar radiation, which would reduce its propagation speed. Also, northerly opposing flow, especially near the surface, tends to increase during periods of high solar radiation, which would slow or stall gravity current propagation. This is consistent with the climatology of Bond et al. (1996), which found that the initiation of wind reversals tended to occur during night or morning.

5. Topographic effects

Figure 5 shows the normal radiation, offset radiation, and no radiation simulations using the realistic terrain shown in Fig. 3b and should be compared with Fig. 4, which shows the same simulations but for the uniform terrain of Fig. 3a. The coast-parallel potential temperature gradient in the normal radiation simulations for both uniform terrain (Fig. 4a) and realistic terrain (Fig. 5a) are somewhat similar in the early phase (i.e., between hours 30 and 42) but differ more obviously in the second phase of evolution.

In the second phase of evolution (i.e., between hours 42 and 84 in Fig. 5a) the translation speed is much less (5.6 m s\(^{-1}\)) in the realistic terrain simulation than in the uniform terrain simulation (9.0 m s\(^{-1}\)). A factor that diminishes the propagation speed in the realistic terrain simulation during phase II is the decreased adiabatic warming as the air flows offshore down the less severe realistic terrain. This will weaken the coast-parallel temperature and pressure gradients and hence lower the propagation speed. Increased friction due to enhanced form drag in the real terrain simulation could also contribute to decreased propagation speed.

By comparing Fig. 4c with Fig. 5c (uniform and realistic terrain runs with no radiation) some indication of the influence of topographic variation in the absence of the diurnal cycle can be inferred. Specifically, a second CTD-like feature forms in the real terrain run (Fig. 5c) at hour 20 and position 1000 km. As there is no real evidence of this feature in the uniform terrain run, its formation is due to the differences between the uniform and realistic terrain and in particular to spatial variations in the lowering of surface coastal pressure via adiabatic warming and suppression of the MBL depth due to local variations in leeside behavior. These ideas are explored further in a subsequent paper (TJR).

6. Discussion and conclusions

SRK’s idealized model of CTDs seems to capture and explain the essential and large-scale evolution of many of the disturbances observed to date. The major features discussed in SRK are also contained in the RAMS simulations presented here: offshore flow preceding and to the north of the disturbance; suppression of the MBL due to the offshore flow; consequent lowering of pressure in the coastal zone, resulting in onshore flow to the south; onshore flow raising the MBL, increasing the pressure, and resulting in the CTD. SRK find nonlinear wave steepening of an initial Kelvin wave–like disturbance into a bore or gravity current occurs shortly after initiation in their idealized simulations as the disturbance accelerates to the north without any diurnal forcing. Since in the SRK model, the MBL is completely evacuated ahead of the CTD, the disturbance in their model must propagate as a gravity current. We also find nonlinear wave steepening but also find a change in the CTD characteristics from borelike to gravity current–like with a resulting change in propagation speed. We find however that the timing and location of this steepening, which we term a transition between phase I and phase II, is a result of the interplay between mesoscale topographic features and diurnal diabatic processes that are not contained in the more idealized modeling approach of SRK. Specifically these features enhance the coastal troughing and therefore the ageostrophic northward acceleration and nonlinear steepening at specific locations and times of day. It is these features that may lead to the sometimes unsteady propagation characteristics reported in observational studies (e.g., Reason and Dunkley 1993).

Unsteady propagation has also been found in the linear shallow-water model results of Rogerson and Samelson (1995). These authors showed that continuous nonresonant periodic synoptic forcing of the marine atmosphere in which a series of synoptic systems continuously cross the coastline can lead to unsteady propagation along the coast of the resulting CTD. However, as there is little observational evidence in cases reported on to date that this type of synoptic forcing is typical of CTDs, it is difficult to accept this mechanism for unsteady propagation as being applicable to CTDs in general and for this case in particular, at this stage of our understanding.

In order to see how a change in dynamical characteristics can account for the change in translation speed between phase I and II, phase speeds for linear and
TABLE 2. Calculated phase speeds based on the RAMS simulation: \( h \) is the unperturbed MBL depth ahead of the CTD, \( h' \) is the perturbed MBL depth behind the CTD, \( \theta_I \) is the potential temperature of the layer above the MBL inversion, \( C_{KW} \) is the linear Kelvin wave phase speed, \( C_{NKW} \) is the nonlinear Kelvin wave phase speed (Reason and Steyn 1992), \( C_m \) is the bore phase speed (Klepp et al. 1997), and \( C_{GC} \) is a 5-h centered average of the gravity current phase speed based upon surface pressure data from RAMS output (Seitter and Muench 1985; Jackson et al. 1999). Here \( g' \), the reduced gravity, is \([\theta_I - \theta]/\theta] \). The heights were estimated based upon the 288-K contour in the coast-parallel vertical cross section. The RAMS observed speed is the translation speed based upon the slope of the locus of maximum coast-parallel potential temperature gradient. Since the flow ahead of the CTD was mostly offshore with little opposing flow, the translation speed is close to the phase speed.

<table>
<thead>
<tr>
<th>Phase</th>
<th>(1200 UTC 16 May)</th>
<th>(0000 UTC 17 May)</th>
</tr>
</thead>
<tbody>
<tr>
<td>( h )</td>
<td>100 m</td>
<td>0 m</td>
</tr>
<tr>
<td>( h' )</td>
<td>500 m</td>
<td>500 m</td>
</tr>
<tr>
<td>( \theta_I )</td>
<td>285 K</td>
<td>287 K</td>
</tr>
<tr>
<td>( \theta_2 )</td>
<td>300 K</td>
<td>302 K</td>
</tr>
<tr>
<td>( C_{KW} = g' h )</td>
<td>7.2 m s(^{-1})</td>
<td>—</td>
</tr>
<tr>
<td>( C_{NKW} = g' h' )</td>
<td>16.1 m s(^{-1})</td>
<td>16.0 m s(^{-1})</td>
</tr>
<tr>
<td>( C_m = \sqrt{(2g/\rho) h'/(h + h')} )</td>
<td>20.7 m s(^{-1})</td>
<td>22.6 m s(^{-1})</td>
</tr>
<tr>
<td>( C_{GC} = 0.79 \Delta p/\rho )</td>
<td>3.9 m s(^{-1})</td>
<td>6.1 m s(^{-1})</td>
</tr>
<tr>
<td>RAMS observed</td>
<td>17.0 m s(^{-1})</td>
<td>5.8 m s(^{-1})</td>
</tr>
</tbody>
</table>

nonlinear Kelvin waves, gravity current, and bore were estimated from the RAMS output for the realistic terrain simulation and compared with one another in both phase I and phase II. The results are summarized in Table 2. In order to estimate the parameters, such as MBL depth and potential temperatures, required to calculate the various phase speeds, it is necessary to approximate or “shoehorn” a continuously stratified atmosphere into a two- or three-layer system. There is necessarily a certain amount of subjectivity in doing this, resulting in a range of possible propagation speeds. We have selected these parameters based upon visual inspection of coast-parallel potential temperature profiles, and estimate that the calculated phase speeds could vary by as much as ±15% from the values we calculate. The 15% error range was found by estimating the largest and smallest reasonable layer depths and applying these to the formulas used to calculate phase speeds. Another uncertainty in calculating theoretical phase speeds with observed (or in this case modeled) translation speeds, is that it is necessary to add the opposing flow onto the observed translation speed to obtain the observed phase speed. Since the flow in the pre-CTD environment for the May 1985 case as well as other cases, tends to be offshore and is also subject to large diurnal variation, it is not obvious how to assess the opposing northerly flow. Figures 6a and 6b show the along-coast wind component between points B and C in Fig. 5 for the uniform and realistic topography. Superimposed upon these plots is shading of the coast-parallel potential temperature gradients from Figs. 4a and 5a. Inspection of Fig. 6b for the realistic topography simulation, reveals that during phase I, at hour 36 (1200 UTC May 16), the flow north of the disturbance leading edge ranges from light favorable southerly flow to 6 m s\(^{-1}\) opposing northerly flow. During phase II, at hour 48 (0000 UTC 17 May), the flow north of the disturbance ranges from 0 to 4 m s\(^{-1}\) northerly. Ralph et al. (2000) were faced with this problem, especially regarding the impact of the diurnal cycle. Their approach was to look at a 1-month composite of wind profiler data below 1.2-km elevation. They determined that the mean northerly flow in this layer was 1.3 ± 0.3 m s\(^{-1}\) during the period 0600 UTC–1700 UTC (2200–0900 LST). The mean northerly flow in this layer is greater during the late afternoon, reaching a maximum of 5.5 m s\(^{-1}\) at 0200 UTC with a daily average of 2.7 m s\(^{-1}\). These values are consistent with the simulation results presented here. Since the simulated translation speeds are calculated during a period of nonpeak solar heating, if one adds a 1.3 m s\(^{-1}\) correction to the RAMS “observed” translation speeds (17.0 and 5.8 m s\(^{-1}\) for phases I and II, respectively) the observed phase speeds are close (certainly near a ±15% uncertainty) to the bore/nonlinear Kelvin wave speed during phase I and the gravity current speed during phase II (Table 2), suggesting that a change in dynamical behavior accounts for the change in translation speed.

Further evidence for this change in dynamical behavior can be found by examining the along-coast winds behind the CTD leading edge in Fig. 6. During phase I of propagation in which the translation speed is approximately 17 m s\(^{-1}\) (i.e., before hour 42 in Fig. 6b), the southerly winds south of the CTD leading edge range from 7 to 15 m s\(^{-1}\), with the stronger winds located 200 km behind the CTD leading edge. During phase II of propagation in which the propagation speed is 5.8 m s\(^{-1}\) (i.e., after hour 42 in Fig. 6b), the southerly winds south of the CTD leading edge range from 6 to 11 m s\(^{-1}\) with winds up to 17 m s\(^{-1}\) 300 or so km behind the CTD leading edge. Since winds in the direction of gravity current propagation must be at least equal to the propagation speed, the modeled CTD in phase I cannot be a gravity current. However wind speed in the direction of propagation being greater than or equal to the propagation speed during phase II is consistent with a gravity current interpretation. During phase II of propagation, the potential temperature structure (Figs. 7e–f) is suggestive of a gravity current interpretation since the isentropes at the disturbance leading edge extend vertically to intersect the surface clearly separating the lighter predisturbance air from the denser postdisturbance air. During phase I of propagation, the potential temperature structure (Figs. 7b,c) is suggestive of a bore or mixed Kelvin wave/bore interpretation. Since a Kelvin wave/bore can propagate faster than the winds behind...
it, this also is consistent with a Kelvin wave/bore interpretation of the CTD during phase I. Ralph et al. (2000) drew a similar conclusion based on observations of the 10–11 June 1994 event except that there was no observed transition to a gravity current—that CTD did not propagate north of Cape Mendocino. Another difference between the interpretation here and that of Ralph et al. (2000) is that the mixed bore/Kelvin wave existed within the MBL, while the present simulations of the May 1985 event suggest that the MBL, even during the Kelvin wave/bore phase, played an active role in the disturbance.

In summary, RAMS simulations suggest that the model of CTD evolution presented in SRK describes well the simulated general evolution of the May 1985 strong CTD event. It is shown that diurnal radiation effects and mesoscale topographic variations, features not modeled in SRK, can significantly modify the CTD dynamical and hence propagation characteristics. Specifically, wave steepening and the transition from bore-like to gravity current-like, with a subsequent change in propagation speed was found to be closely tied to the diurnal cycle and its interplay with mesoscale topographic variation.

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