The Initiation and Horizontal Scale Selection of Convection over Gently Sloping Terrain

JEAN-LUC REDELSPERGER* AND TERRY L. CLARK

National Center for Atmospheric Research,** Boulder, Colorado

(Manuscript received 9 February 1989, in final form 29 September 1989)

ABSTRACT

Two- and three-dimensional numerical simulations were performed to investigate the scale selection and initiation of both moist and dry convection over gentle western and gentle eastern slopes where the latter represents an idealization of the eastern Colorado region of the Great Plains of North America. This work extends earlier studies of thermally forced convection by considering a model framework that is large enough to resolve both the convective scale dynamics as well as the larger scale dynamics within which the convection is embedded. As a result, the scale interaction problem leading to the selection of the dominant deep modes of the troposphere and consequent convection initiation is more realistically treated. The main physical mechanisms involved in the initiation of convection in these studies are the usual boundary-layer instabilities leading to the development of eddies and/or shear-aligned rolls, the excitation of gravity waves by the boundary-layer motions interacting with the free atmosphere, and the eventual development of coherent vertical structures that link the boundary layer motions and the overlying gravity waves into larger horizontally spaced modes than typically obtained from an isolated boundary layer.

It has previously been shown that the mean wind shear spanning the region between the top of the boundary layer and the overlying stable layer plays an important role in producing energetic deep modes in the presence of thermal forcing. In the present simulation this shear results from a combination of initial baroclinicity associated with the westerlies and production by the differential thermal gradients formed by heating gently sloping terrain. Westerly geostrophic shears of either 3 or 5 m s⁻¹ km⁻¹ over the first 3.5 km above sea level were used as initial conditions. A balance is maintained between shear production through large-scale forcing and shear destruction through boundary-layer mixing that results in significant shear. The experiments showed a broad range of responses as a consequence of the horizontal variability of the shear structures. The preferred region of both dry and moist convection was found to be the eastern slope where the terrain effects result in an enhancement of the low-level shear. In response to the directional structure of the shear spanning the boundary layer and free atmosphere both a banded and a less coherent scattered organization were obtained for the waves and clouds.

Dominant deep modes were found to organize and initiate moist convection. West–east horizontal scales of the deep modes in the dry experiments were found to range from about 11 to 28 km with either a banded or a cellular structure with scales between 4 to 6 km in the south–north direction. The timing of the onset of the moist convection appeared to affect the final horizontal-scale selection in the moist experiments. The moist convection appears to lock onto the scales of the dry modes that initiate the convection for these particular experiments. The largest horizontal scales of dominant modes in the dry experiments were about 28 km and developed rather slowly as compared with the 11 km scale dominant modes. These largest horizontal scales did not develop in the moist experiments where clouds appeared early but did develop in those moist experiments where moist convection took longer to develop.

1. Introduction

Observational studies have shown that the formation of storms is generally associated with organized lifting of low-level moist air at horizontal scales ranging from 10 to 20 km (e.g., Fankhauser et al. 1982; Wilson and Schreiber 1986; Wilson 1986; Brady and Szoke 1989). Large-scale disturbances such as fronts, dry lines or topography can provide such lifting and initiate moist convection when the atmosphere is conditionally unstable. However, deep convection is also observed to develop in the absence of such large-scale forcing. In this case, we must consider local initiation processes involving the interaction of the boundary layer and the overlying troposphere.

Recently, Clark et al. (1986; hereafter CHK), Balaji and Clark (1988; hereafter BC), and Hauf and Clark (1988; hereafter HC) have presented evidence of such processes through numerical simulations at the convective scale (using horizontal domain sizes from 30 to 60 km). The problem of convection initiation was

* Permanent affiliation: Centre National de Recherche Météorologique, CNRS and DMM, Toulouse, France.
** The National Center for Atmospheric Research is sponsored by the National Science Foundation.

Corresponding author address: Dr. Terry L. Clark, NCAR/MMM,
P.O. Box 3000, Boulder, CO 80307-3000.

© 1990 American Meteorological Society
studied in terms of the atmospheric response to surface sensible heat flux. These studies concentrated on cases having significant levels of mean-wind shear in the region spanning the top of the boundary layer and the lower levels of the overlying stable layer. It was shown that the excitation of gravity waves in the free atmosphere overlying the boundary layer represents a possible mechanism for generating the initial lifting at these observed scales. The wave excitation was produced by a Reynolds-stress effect where boundary-layer eddies act as small hills in the presence of mean wind shear, as suggested by Haman (1962) and Mason and Sykes (1982) in the interpretation of their numerical simulations. The requirement that some of the air must flow over the eddies results in the excitation of gravity waves. Using such an excitation mechanism, CHK, BC, and HC obtained a horizontal scale selection of deep modes between about 6 and 12 km that is in reasonable agreement with those observed in nature. These studies emphasized the importance of the alignment of the boundary layer and overlying free-tropospheric shear in determining the character of the deep modal response. From this point of view, the HC and BC cases represent two extrema. Indeed, the directional shear of the BC case resulted in longitudinal boundary-layer rolls aligned parallel with bands of gravity waves in the overlying stable layer, whereas the unidirectional shear of the HC case resulted in fully three-dimensional waves and eddies. In this latter case longitudinal rolls first form in the boundary layer but as soon as gravity waves are excited the wave fronts and boundary-layer eddies interfere because gravity wave fronts tend to form perpendicular to the wind shear direction (e.g., Gossard and Hooke 1975). This major difference of preferred orientation resulted in a scattered (or three-dimensional) global structure for both the eddies in the boundary layer and gravity waves aloft in the HC simulations.

One of the shortcomings of these previous works is that it was necessary to artificially maintain this spanning shear through horizontal pressure gradients (HC) or using an unrealistic value of surface friction (BC) during the simulations. It is well known that the momentum mixing due to boundary-layer eddies destroys the shear over a short time scale (e.g., Wyngaard et al. 1974). Using such artificial procedures of maintaining shear, these previous studies demonstrated that small scale convective processes were able to initiate deep convection in otherwise undisturbed flow by excitation of deep normal modes. The relative importance of these ad hoc procedures can only be evaluated through simulations where both the small and large-scale effects are included. Another shortcoming is the relatively small horizontal cyclic domain size (60 km) used in the previous studies. Cyclic lateral-boundary conditions were necessary to maintain near field forcing conditions during the 2–3 hours of the initiation process. The simulation of large scales primarily concerns the processes governing the creation of such shear and maintaining it against the momentum mixing in the boundary layer on convective time scales.

The present study extends the results of BC and HC by directly simulating both the small convective and larger scale effects in the same framework using interactive grid nesting techniques. The mechanism of maintaining shear in the boundary layer is through the differential heating of air above a gentle slope that represents an idealization of the eastern slope of the Great Plains. Section 2 outlines the methodology, including the linear and nonlinear models used in this study and the initialization procedure. The mesoscale results are presented in section 3 with particular emphasis on the shear structure in the boundary layer. Section 4 deals with two- and three-dimensional dry simulations using two different initial baroclinicities. In section 5, results are presented from three-dimensional simulations using the complete bulk warm-rain physics parameterization of the model.

2. Methodology

a. Numerical models

Nonlinear and linear models were used in the present study of the initiation of deep convection on a gentle slope. In order to simulate the range in scales necessary to represent both the convective motions as well as the larger scale dynamics associated with the gentle slope, interactive grid nesting techniques were employed. The nonlinear and nonhydrostatic anelastic model of Clark (1977, 1979), which was further developed to include grid nesting by Clark and Farley (1984), is used in the present study. This model now uses Smolarkiewicz’s multidimensional positive-definite advective transport algorithm for all thermodynamic fields as described in Smolarkiewicz (1984) and Smolarkiewicz and Clark (1986).

A linear spectral model, as described in BC, is also used in this study to distinguish dry linear from nonlinear effects in the development of the deep modes. This model solves the dry-physical linearized equations of the nonlinear model as an initial value problem and is used to provide a modal analysis of a deep atmosphere. The mean vertical profiles of wind and potential temperature are arbitrary functions of height resulting in a nonseparable form to the equations in the vertical dimension. The modal response of the isolated boundary layer forms only part of the solution and is located at relatively high horizontal wavenumbers. There are also low wavenumber solutions representing the deep modes of interest in the present study that are comprised of boundary-layer eddies and overlying gravity waves.

Since both models are already described in the literature, we refer interested readers to the cited papers.
for further details. It was, however, necessary to design a new radiation scheme for the model that accurately predicts the longwave radiative cooling throughout the troposphere. Previous studies with Clark’s model had used the radiative scheme of Sommeria (1976) which was designed for boundary-layer studies. Application of this scheme to the full atmosphere for diurnal cycles resulted in erroneous cooling and heating rates in the nocturnal free atmosphere. An accurate representation of the radiative processes in a three-dimensional cloud model should account for cloud effects and yet be computationally efficient. Two main approximations have been proposed in the literature to simplify the computation of the radiative-transfer integral. Sasamori (1972) assumed an isothermal atmosphere. In a boundary-layer study, Sommeria (1976) neglected the radiative exchanges between the atmospheric layers. In the present case, both of these approximations failed to produce realistic results. A method suggested for a three-dimensional boundary-layer model by Veyre et al. (1980) has been modified and generalized. The results produced by the scheme are close to the full computation (using double vertical integrals) and the scheme is computationally affordable. This new scheme accounts for carbon dioxide and water vapor as well as liquid water and is described in the Appendix.

b. Outline of experiments

Rather than attempt to simulate a real case, idealized initial conditions were used to gain more insight into the processes of convective initiation. The topography chosen is an idealized representation of the Great Plains. From both western and eastern sides of a 200 km wide plateau, a gentle slope of 1/400 is specified. The projected length of the slopes is 400 km resulting in a plateau height of 1000 m. The domain size is 2400 × 20 km in the x (west–east) and z directions, respectively. A 20 km cross-stream (south–north) dimension is added for the three-dimensional experiments. Cyclic x and y boundary conditions are used for the outer domain. In the text, the 100–500 km and 700–1100 km regions will be referred to as the western and eastern slopes, respectively. The initial temperature field corresponds to two horizontally homogeneous layers of constant stability, dlnθ/dz = 10−3 m−1 from z = 0 to 10 km and 2 × 10−5 m−1 above 10 km, respectively. A surface temperature of 300 K and pressure of 1000 mb was specified at z = 0. Westerlies are prescribed, associated with global-scale baroclinicity, having constant shear through a depth of 5.5 km (5 × 10−3 s−1 for experiments MESA, DRY2A, DRY3A, and WETA, and 3 × 10−3 s−1 for experiments MEB, DRY2B, DRY3B, and WETB. The water vapor mixing ratio is specified as linearly decreasing with height from qv = 11.75 at z = 0 to qv = 1.5 g kg−1 at z = 4.75 km. From z = 4.75 to 7.25 km qv logarithmically decays to 0.1 g kg−1 and above z = 7.25 km qv logarithmically decays at −0.563 g kg−1 km−1. The model was first run in its two-dimensional version from this initial state for 6 h, beginning at 0600 LST (sunrise). A spatially uniform surface heat flux varied sinusoidally from zero at 0600 LST to its maximum value of 250 W m−2 at 1200 LST. Surface momentum fluxes are specified through a bulk drag formulation using a drag coefficient of 2 × 10−3 s−1 in the basic experiments.

Two early experiments, MESA and MEB, were performed using the two-dimensional dry version of the model (see Table 1 for details). In the dry version, all moist processes are turned off except for the advection of the water vapor field used to determine the radiative cooling. These experiments were designed to show the basic physical mesoscale response and used mesh sizes of 6 and 0.5 km in the x and z directions, respectively. Other experiments, not included in the table, with different surface sensible heat flux and drag-coefficient amplitude will also be briefly discussed.

Table 1 also shows six further experiments that were performed using the two-way interactive-nesting procedure. The purpose of these experiments is to interactively simulate both the mesoscale features (as in experiments MESA and MEB) and the convective scale processes occurring on the sloping terrain. In all of these experiments the outer model has the same initial characteristics. The inner model corresponds to a domain of 1200 km horizontally and 12 km vertically, centered on the plateau. The resolution for the inner model is 2 and 0.25 km in the x and z directions, respectively. Experiments DRY2A and DRY2B are dry physics, two-dimensional experiments whereas experiments DRY3A and DRY3B are three-dimensional using a resolution of 1 km along the y-direction. This nonisotropic horizontal resolution was chosen based upon three-dimensional linear model simulations, as well as previous studies of HC and BC. Indeed, these results show that the horizontal scale in the shear direction of stable layer is typically about twice that in the cross-stream direction. Thus, similar resolution in both directions can be assumed for the scattered gravity-wave response regions. However, this choice may possibly be poor to accurately represent the internal cloud dynamics; that is beyond the focus of this paper. Experiments WETA and WETB are the same as DRY3A and DRY3B, except they consider the full warm rain physics of the model. All the three-dimensional simulations are initialized using the “dry” two-dimensional simulations having the identical large-scale mean flow. The two-dimensional fields were uniformly extended in the north–south direction and the warm rain physics was turned on in the case of the moist experiments using the existing qv fields from the dry experiments. In order to obtain initial three-dimensional perturbations, a 5 percent random perturbation (∼12 W m−2) on the surface sensible heat flux
is applied along the south–north direction during the first time step (at 1130 LST) of the three-dimensional experiments.

3. Mesoscale wind structure

One possible process that is able to maintain directional shear is the low-level jet of the United States Great Plains. Southerly boundary layer winds blow normal to the westerly winds aloft producing such directional shear. Generally, these low-level wind maxima are considered to be related to thermal forcing, although other hypotheses have been given (Wexler 1961; Uccellini and Johnson 1979). There are two principal mechanisms proposed to explain how the thermal forcing contributes to the jets' formation. These involve first, the time variation of the eddy viscosity in the boundary layer (Blackadar 1957; Buajjiti and Blackadar 1957) and second, the creation of a diurnal thermal wind over sloping terrain (Bleecker and André 1951; Holton 1968).

It now seems well established that the effect of the gentle sloping terrain over the Great Plains is the major physical factor involved in these phenomena (Lettau 1967; Bonner 1968; Bonner and Paegle 1970; McNider and Pielke 1981). The driving mechanism is the diurnal heating and cooling cycle above the slope, resulting in temperature gradients on horizontal surfaces above the slope. The effects of surface friction, eddy mixing, Coriolis forces, and advection are also important in the force balance. As previously discussed in the literature, low-level jets occur both day and night with a maximum during the night.

Although the formation of the low-level jet has been thoroughly studied in the literature, its role has received little attention. Previous studies of convection initiation have shown that the mean wind shear in the region spanning the boundary layer and free atmosphere has an important coupling effect. The low-level jet is a natural mechanism providing such coupling possibly throughout a diurnal period. The present interest lies in studying the daytime influence of this jet and its role in initiating tropospheric convection in the near field of thermal forcing. The effect of the nocturnal jet is beyond the scope of this paper even though its influence still relates to a layer coupling but in a far-field sense where the source of kinetic energy for initiating the convection is more likely found in the free atmosphere as a result of previous daytime activity. The shear obtained through such differential daytime heating effects of the sloping terrain will now be described.

Figure 1 shows a shear zone, with a south wind maximum over the eastern part of the eastern slope produced by the horizontal thermal-gradient forcing. As shown in numerical simulations by McNider and Pielke (1981), the combination of the frictional stress and the thermal-wind shear leads to a broad maximum in the south–north wind component near the top of the boundary layer. Without mean flow one would expect an antisymmetrical distribution of north–south wind structures about the center of the plateau. However, inclusion of mean flow advection distorts this pattern as seen in Fig. 1a. More interesting in our case is the creation of shear in the region spanning the boundary-layer top and overlying layer because this shear plays a major role in defining the character of the deep modes as shown by CHK, BC, and HC. The physical processes creating such shear are important in maintaining a portion of the shear against the effect of mixing by the boundary-layer eddies. Figures 1a and 1b clearly indicate stronger shear along the eastern slope than the other regions for experiment MESA. This is also the case for experiment MESB as well as the other experiments described in Table 1. These differences
between eastern and western slope shears are attributable to the fact that the thermal upslope wind is in opposition to the westerly wind on the eastern slope and as a result adds low-level shear, whereas the opposite occurs on the western slope. Furthermore, the vertical variation of the shear is stronger on the eastern part of the eastern slope when compared to its western part. Previous findings suggest that the deep modal development should be stronger on the eastern slope where the spanning shear is stronger, even though earlier-time boundary-layer eddy formation is likely to occur in the regions of weaker shear.

Figures 2a–d show the magnitude and north-relative direction of the shear for experiments MESA and MESB at 1500 LST. The shear magnitude and direction have similar overall horizontal and vertical structures in the two experiments where the direction turns by as much as 75 deg between the middle of the boundary layer and the overlying stable layer. This is less than the 90 deg turning in the three-dimensional BC case that resulted in an overall banded structure for both eddies and waves. However, MESA and MESB are two-dimensional experiments and accordingly affected by this assumption. The fields of shear that occur in the three-dimensional experiments are similar in structure to those shown in Fig. 2 but with regions of even greater directional turning. The vertical variation of the shear direction is more evenly distributed along the eastern slope in experiment MESB than in MESA at this time. The shear is generally more directional in the eastern part of the eastern slope for both experiments and even more so in the MESB than in MESA. This is a result of the amplitude of the south–north shear, due to the differential thermal gradient forcing, being proportionally larger in MESB than in MESA relative to the initial west–east shear. Maximum local-shear magnitudes of $8 \times 10^{-2}$ s$^{-1}$ occur at the eastern and western part of the eastern slope and at heights of 1.5 and 1 km above the ground for MESA (Fig. 2a). Maximum local shear magnitudes of $6 \times 10^{-3}$ s$^{-1}$ occur for MESB with a similar distribution on the eastern slope (Fig. 2c). All these differences between the two experiments have important consequences on the behavior of the convective initiation along the eastern slope.

Some sensitivities of the locally produced shear structure to initial parameters, besides the initial baroclinicity, were also studied. Experiments with a constant initial easterly wind of 15 m s$^{-1}$ and no initial baroclinicity were performed to study the effect of drag coefficient and surface sensible heat-flux maxima. The main effect of increasing the drag coefficient is to increase the magnitude of the shear with negligible effect on its direction. At noon, the shear maximum varies from $3.4 \times 10^{-3}$ s$^{-1}$ to $27 \times 10^{-3}$ s$^{-1}$, for a variation of drag coefficient from 0 to $10^{-2}$ s$^{-1}$. The jet height also changes from 0 to 2 km above the ground, for the same variation of drag coefficient. The sensitivity to changes in the surface heat flux is a little more complicated. An increase of the surface heat flux magnitude results in a decrease of the absolute shear maximum. However, the height of the shear layer changes more or less in proportion to the boundary-layer height. Sensitivity experiments were not performed with background shear included and we do not know if a similar result would have been obtained. However, the present tests suggest that it is important to accurately treat the surface friction and the sensible heat flux.

4. The dry convective development

a. Two-dimensional case

Both linear analysis of unstably stratified boundary-layer instabilities and numerical simulations show that the most unstable boundary-layer modes are longitudinal rolls aligned with the mean shear. These modes cannot appear in the two-dimensional simulations DRY2A and DRY2B and the simulated instabilities therefore will represent spurious oblique modes; however, the relative simplicity of the two-dimensional simulations warrants their consideration for designing three-dimensional cases. Gravity waves can be modeled in two dimensions in the plane of the mean flow and the sensitivity of their final structure to the tropospheric characteristics can be studied as in CHK. Also, the two-dimensional simulations were used to test the new radiation scheme employed in the model for the flow integration periods. The principal shortcoming is that
the dominant deep mode of the two-dimensional case may not be that which is obtained in three dimensions when the full effect of shear between the boundary layer and overlying layer is considered.

Before describing the development of the dominant forced deep mode, it is worth noticing the large difference between the response using a horizontal mesh size of 6 km for experiment MESA (Fig. 3a) and the one for the present experiment DRY2A using a horizontal mesh size of 2 km (Fig. 3b). Note that in Fig. 3, MESA is purposely shown at a much later time to allow for any late onset of convection. Its absence minimizes the possibility of the differences being simply due to timing. The principal reason for this large difference is that simulation MESA is not able to resolve the boundary layer eddies. This result may have consequences for mesoscale or regional scale models. Even with a good representation of moist processes (e.g., including equations for cloud and rainwater content) it is impossible to simulate such convective initiation processes unless sufficient horizontal resolution is used to “turn on” the appropriate physics. Moreover, as will be dis-

![Figure 2](http://journals.ametsoc.org/jas/article-pdf/47/4/516/3425467/1520-0469(1990)047_0516_tiahss_2_0_co_2.pdf)

**Fig. 2.** Vertical cross section of shear at 1500 LST: (a) and (b) show the shear magnitude (contour interval 1 m s\(^{-1}\) km\(^{-1}\)) and direction relative to north (contour interval 15 degrees), respectively, for experiment MESA. Similarly, (c) and (d) are for experiment MESB using the same contour intervals.
cussed in section 4c, the Reynolds stress associated with these resolved waves may be significant.

The magnitude of the shear spanning the region between the top of the boundary layer and the overlying stable layer is considered important in determining the character of the deep modes. As already discussed, there are significant horizontal variations in the character of the shear for the present cases (Fig. 2). Consequently, a commensurate variability in modal response throughout the domain is expected. The horizontal variation of the shear also leads to further difficulties in analysis because of the different temporal rates of evolution throughout the domain. Figure 3b presents a plot of the vertical velocity field for the inner domain of experiment DRY2A at 1200 LST. This is roughly the time when eddies have just filled the boundary layer in some regions and the first gravity waves are beginning to appear as shown in a reduced section in Fig. 4a. In experiment DRY2A, the boundary-layer eddies are not as well developed at 1200 LST for the 800–1100 km section on the eastern slope as in the 500–800 km section (Fig. 3b). This may be a consequence of the large magnitude of shear in the boundary layer inhibiting roll development, as discussed by Asai (1970, 1972) and Kuettner (1971) and by CHK. Since the deep modal response is significantly weaker for the western than the eastern slope, only results for the eastern slope will subsequently be shown in this section. As previously discussed this difference is attributed to the large differences of the shear magnitude in the region spanning the top of the boundary layer (Fig. 1). Further comparisons between eastern and western slopes will be discussed for the three-dimensional case in section 4b.

Figures 4a–d show four time levels of the vertical velocity, w, for experiment DRY2A for the inner domain. As described in the Introduction, the shallow Rayleigh modes first develop in the boundary layer in areas of minimum shear (Fig. 4a). Once the eddies have filled the boundary layer they act as obstacles to the flow above the boundary layer. As a consequence of the blocking effect of these eddies, internal gravity waves are excited and propagate vertically into the stable troposphere. This effect is possibly overestimated in the two-dimensional case, as the flow is not able to flow around the eddies. The boundary-layer eddies gradually develop along the slope and at 1240 LST are present everywhere. The deep modes develop with the horizontal scales gradually increasing with time (Figs. 4b, 4c, and 4d). The horizontal wavelength of the dominant deep mode (i.e., the deep mode having the maximum kinetic energy) ranges from about 11 km at noon to about 23 km at 1500 LST. The mean flow field is rather inhomogeneous along the slope, contrary to the previous studies, resulting in detectable differences in modal response. Note also how the horizontal scale of the boundary-layer eddies closely follows that of the gravity waves, resulting in a nearly uniform hor-

![Figure 3](http://journals.ametsoc.org/images/jas/47/4/522_f03_jas4704_0014.jpg)

**FIG. 3.** Vertical cross sections of w from (a) MESA at 1500 LST and (b) DRY2A at 1200 LST. Only the x = 0 to 1200 km part of the domain is shown, which uses 6 and 2 km horizontal resolution in (a) and (b), respectively. Dashed contours show negative values. Note the 20 to 1 ratio in contour intervals between (a) and (b).
modes between the experiments because of the transient nature of the gravity waves and eddies. However, the amplitude of the gravity waves was found to be weaker in DRY2B. The amplitude of $w$ in the gravity waves seems to vary proportionally with the shear magnitude, similar to the results obtained by CHK. The horizontal scale selection of the deep modes is quite similar between DRY2B and DRY2A. However, in DRY2B at 1500 LST there is evidence of a much stronger upstream tilt to the phase lines of $w$ with height suggesting an enhanced vertical propagation of wave energy. This increase in intensity is probably affected by the two-dimensional framework, which does not allow for the decorrelation between the boundary-layer eddies and overlying gravity waves that can occur in three dimensions.

**b. Three-dimensional case**

As discussed in the introduction, the alignment of the boundary-layer and overlying free-atmospheric shear is fundamental in determining the deep mode structure. In the present case, the angle between shear direction in the middle of the boundary layer and the overlying layer varies from about zero to 70 degrees for the two-dimensional case (Fig. 2b) and even stronger directional shear occurs in the three-dimensional cases. Consequently, we should expect a large range of responses throughout the domain with a resulting structure ranging from the HC to the BC case. A larger range of responses may also be expected in three dimensions than in the two-dimensional case because of the fundamental importance of the directional shear in the latter.

![Cross sections of w for experiment DRY2A](image)

**Fig. 4.** Cross sections of $w$ for experiment DRY2A over the eastern slope of the domain. The times shown are (a) 1200 LST, (b) 1300 LST, (c) 1400 LST, and (d) 1500 LST. The contour interval is 0.1 m s$^{-1}$ in (a) and 0.3 m s$^{-1}$ in (b)–(d).
1) Experiment DRY3A

Three-dimensional perspective views of positive $w$ are shown at 1440 LST for experiment DRY3A, for the eastern slope in the boundary layer (Fig. 6c), the stable layer (Fig. 6d), and for the western slope in the stable layer (Fig. 7). These figures give an indication of the variability in response to similar thermodynamic and geostrophic conditions but with the shear structures in the first few kilometers of the atmosphere modified by the gentle slope. As expected from the initial shear structure and from the results of the two-dimensional experiments, the gravity waves are more active on the eastern than the western slope.

Power spectra of $w$ computed for each slope are shown in Figs. 8a and 8b where cyclicity was assumed for the perturbation field over the 400 km extent of the slope. These spectra represent one-dimensional spectra in terms of wavenumber component $k_x$, where an integration was performed over $y$ and $z$. The western-slope spectrum has about half the energy level of the eastern slope. At 1310 LST (Figs. 6a and 6b), gravity waves have already appeared on the eastern part of the eastern slope and are seen to be connected to the boundary-layer eddies. At this time, the initial longitudinal rolls in the boundary layer (forming along the east-west direction) are broken. As discussed in HC and in the beginning of this subsection, this rather scattered structure is a direct consequence of the incompatibility between the preferred boundary-layer eddy and gravity-wave orientation. Their relative orientations are directly related to the angle between the shear in the boundary layer and the overlying free atmosphere. Three-dimensional structures are also ob-

---

![Fig. 5. Same as Fig. 4 except for experiment DRY2B.](image-url)
tained for the gravity waves. The spectral power of the vertical velocity at 1310 LST indicates mean horizontal wavelengths of about 11–13 km along the west–east direction (Fig. 8b) and 4 km along the south–north direction (not shown) ($k_x = 31, 37$ and $k_y = 5$). Power spectra limited to the boundary layer as well as that limited to the stable layer (both not shown) show similar energy levels and modal wavelength peaks. The temporal evolution indicates that the deep modes are still growing. At later times, the deep modes gradually

**Fig. 6.** Perspective views of $w$ on the eastern slope for experiment DRY3A. Surfaces shown are for $w > 0.5$ m s$^{-1}$; (a) and (b) show the boundary layer ($z < 2$ km above ground level) and the stable layer ($z > 2$ km above ground level) at 1310 LST, respectively. Similarly, (c) and (d) are for 1440 LST.
gain in amplitude and exceed that of the shallow Rayleigh modes.

Figures 6c and 6d show a more mature stage of deep modes at 1440 LST at which time the deep modes completely dominate the boundary-layer eddies and show a selection of two distinct horizontal scales of 11–13 km \((k_x = 36 \text{ and } 31)\) and 28.6 km \((k_x = 14)\) along the east–west direction (Fig. 8b). The latter corresponds to the wave packets, localized in the eastern side around \(x = 1000\) km. This larger horizontal scale is selected later than the smaller scales. This point is considered important to the moist experiments, because the nonlinear effects due to deep clouds can strongly influence the selection of the later time, large horizontal-scale deep modes.

It is worth noting that the deep-mode organization along the south–north direction varies along the slope. This is a direct consequence of the directional shear variations that have been shown in previous simple studies of BC and HC to be a determining parameter. In the present case, the convective organization ranges from banded to scattered regions. The gravity waves are sufficiently intense to modify the horizontal scale of the boundary-layer eddies to scales similar to those of the waves. The selected scale along the south–north direction now varies from 6 to 20 km, but could be different with a larger domain and similar resolution in this direction as the present spectra for the cross-slope direction are based on only 20 points.

Dynamical activity of a wave activity maximum at the eastern part of the slope may be easily explained as a consequence of a vertical variation of shear direction of about 90 degrees in the upper regions of the boundary layer. Indeed this region shows overall structures similar to the pure directional shear of BC.

To get more insight in the processes involved in the deep mode development, the discussion now focuses on the most active region of gravity-wave development that was located in the 800–900 km part of the eastern slope of DRY3A. The times of 1330, 1500, and 1510 LST have been chosen, as they correspond to three different stages for the deep modes in this part of the domain (Figs. 9a–f). At 1330 LST the deep modes are still growing and most of the energy is still restricted to the boundary layer. The eddies have filled the boundary layer and are protruding through a shear layer into the lower part of the stable layer. The dominant mode waveforms found by spectral analysis at this time are \(\lambda_x = 11 \text{ km } (k_x = 9)\) and \(\lambda_y = 4 \text{ km } (k_y = 5)\), in the west–east and south–north directions, respectively (Fig. 10a). At 1500 LST, the gravity waves have shifted to a longer horizontal wavelength resulting in more extended boundary-layer eddy structures. The comparison of Figs. 9b and 9d corresponding to perspective views of the boundary-layer motions at 1330 and 1500 LST shows this dramatic change. The dominant wavelengths in the boundary layer are now \(\lambda_x = 10 \text{ to } 20 \text{ km}\), and \(\lambda_y = 5 \text{ to } 6 \text{ km}\) at 1500 LST. The

**Fig. 7.** Perspective view of \(w\) on the western slope stable layer (\(z > 2\) km above ground level) for experiment DRY3A at 1440 LST. Surface shown are for \(w > 0.5\) m s\(^{-1}\).
The structure of eddies in the boundary layer is quite similar but with longer west–east bands of $w$ having a maximum horizontal extent of about 40 km (Fig. 11a) on the 700–800 km section of the eastern slope. Smaller variations of the shear magnitude in the boundary layer account for this difference between the two experiments (Fig. 2d). However, the differences in mean-shear magnitude and direction near the top of the boundary layer appear to have changed the deep-mode organization. During the full simulation period, the gravity waves are clearly more structured in south–north bands for experiment DRY3B than for DRY3A. One large and active band can be observed around 1000 km (Fig. 11b). Two other less intense bands also occur on the western section of the eastern slope. This banded character is related to the structure of the shear direction in this region. Its vertical structure is very close to a pure directional shear case, similar to BC, where the

Fig. 8. One-dimensional power spectra of $w$ along the west–east direction integrated over the full vertical extent of the inner domain. Here $k = 1$ refers to the $x$-direction wavelength of 400 km. The heavy and light solid lines refer to $t = 1310$ LST and $t = 1440$ LST, respectively. The numbers represent wavenumbers; (a) and (b) represent the western slope and eastern slope for experiment DRY3A, respectively; and (c) and (d) represent the western slope and eastern slope for experiment DRY3B, respectively.
veer is nearly 90° across the boundary layer in these banded regions. Another factor is the smaller speed difference between the two layers. The presence of the directional shear is assumed to have reduced any effective interference between the banded gravity wave structure and the boundary layer eddies as compared to DRY3A. The net result is that DRY3B has only slightly less energy in the gravity waves than DRY3A. As in the two-dimensional case, the magnitude of $w$ in the gravity waves is found to be weaker in DRY3B than in DRY3A. The power spectrum of $w$ for western and eastern slopes (Figs. 8c and 8d, respectively) shows slightly less energy in DRY3B than for DRY3A. The selected horizontal wavelengths along the west–east direction are also slightly different with a larger range of deep-mode scales: $\lambda = 11, 17.4, 19, 26.7$ km ($k_x = 35, 23, 21, \text{and } 15$).

3) LINEAR EXPERIMENT

In order to distinguish between linear and nonlinear processes, the linear model discussed in section 2a was run using the wind and thermodynamic profiles from the nonlinear simulation DRY3A averaged over the 800–900 km region. A domain of $60 \times 60 \times 16$ km using the same grid size as in the nonlinear model was chosen. All sloping terrain effects, however, are excluded from the linear model. Figure 12 presents one-dimensional power spectra of $w$ taken along the west–east direction after 40 and 60 min simulation time. The deep modes begin to develop after about 40 min at which time two modes are present, corresponding to wavelengths of 10 and $20$ km ($k_x = 6$ and 3). These wavelengths, as well as those in the other direction, are close to those in the nonlinear simulations prior to the

---

**Fig. 9.** Perspective views of surfaces of $w > 0.5$ m s$^{-1}$ for the 800–900 km section of the eastern slope from experiment DRY3A: (a), (c) and (e) show the stable layer ($z > 2$ km above ground level) at 1330, 1500 and 1510 LST, respectively; (b), (d), and (f) show the same respective times for the boundary layer ($z \leq 2$ km above ground level).
development of a large convective packet. After 60 min of simulation time, only one wavelength (20 km) is selected by the linear model in contrast with the nonlinear results where different wavelengths continue to coexist during the later stages of the simulation. There are a number of possible reasons for these differences. Nonlinear wave–wave interactions or horizontal variations of the mean state are two possible causes. One would have to consider horizontally inhomogeneous shear resulting in variable coefficients for the linear model in order to relax the assumption of local horizontal homogeneity. The assumption of cyclicity is also rather questionable for these cases. No attempt was made to account for the horizontal trends. The linear model results, however, indicate that deep modes of similar horizontal scales may exist as natural frequencies of the dry state but the detailed comparison of the spectra is weak. The ground-relative horizontal phase velocity of the gravity waves is close to the wind velocity at the level just above the boundary layer in both experiments. The gravity waves appear to be locked onto the boundary-layer eddies that provide their energy.

c. Reynolds stress

An important effect of gravity waves is the vertical transport of mean horizontal momentum. It has been recently suggested that gravity waves play a role in compensating Rossby wave momentum flux in the upper troposphere (Boer et al. 1984; Palmer et al. 1986). Bretherton (1969) and Lilly (1972) were among the first to point out that gravity waves generated by orography may be important to the general circulation. It has been argued that the neglect of downward momentum flux due to mountain gravity waves may be responsible in part for the systematic overestimation of the westerlies in general circulation models.

Considering that large areas may have convective activity favorable to the development of convective waves, it is conceivable that convective waves may also contribute to the momentum budget. Although mountain-drag estimates are useful in determining relative global significance of drag effect, convective waves are more similar to “cumulus friction” (Schneider and Lindzen 1976), which vertically redistributes momentum and has no net drag effect. It is of interest to examine such effects in the simulations. Figures 13a–f show the profiles of the resolved Reynolds stress components along the west–east direction \( \tau_{xx} \) and along the south–north direction \( \tau_y \) at 1300, 1330, 1400, and 1430 LST, for experiments DRY3A, DRY3B, and DRY2A. The profiles correspond to an average along the full south–north direction but only for the eastern slope (700–1100 km) region. The height of maximum momentum flux divergence is found in the boundary layer for \( \tau_x \). Momentum flux divergence in the overlying stable layer is largest for the strong shear experiments DRY3A and DRY2A. The \( \tau_y \) component is generally about one order of magnitude smaller than

![Fig. 10](http://journals.ametsoc.org/jas/article-pdf/47/4/516/3425467/1520-0469(1990)047_0516_tiahss_2_0_co_2.pdf)  
**Fig. 10.** One-dimensional power spectra of \( w \) for the 800–900 km portion of the eastern slope for experiment DRY3A. Here \( k = 1 \) refers to the x-direction wavelength of 100 km. Times shown are (a) 1330 LST, (b) 1500 LST, and (c) 1510 LST.
\( \tau_x \). To estimate physical significance, the present amplitude of the west–east component (0.2–0.3 kg m\(^{-1}\) s\(^{-2}\)) occurring on the 400 km eastern slope section may be compared to typical values estimated for squall-line mean momentum transport. For example, Laffen et al. (1988) performed a complete momentum budget for a tropical squall line using both Doppler-radar and simulation data. In contrast to the present study, a pressure gradient balanced a large part of the momentum divergence. They estimated their vertical momentum flux to be 1.5 kg m\(^{-1}\) s\(^{-2}\) over an equivalent 400 km domain extent resulting in a wind speed change of about 5 m s\(^{-1}\). In the present case, the maximum rate of momentum transport due to \( \tau_x \) in Fig. 13a amounts to about a 5–6 m s\(^{-1}\) change between \( z = 1.5 \) and 2.5 km for air moving through the eastern slope region at a speed of 15 m s\(^{-1}\). Complicating the comparison, however, are fundamental differences between the phenomena. In summary, the present results suggest that convectively forced gravity waves may significantly contribute to the momentum loss of the zonal wind. Further observational and numerical studies are necessary to assess these preliminary findings.

5. Development of deep convection

The results from the dry simulations have shown that dominant deep modes can be created with horizontal scales similar to those that may trigger observed deep convection. These modes should be capable of triggering deep convective clouds given sufficient moisture. Nonlinearities due to latent heating may, however, strongly affect the scale selection, although BC found that this was not necessarily the case, at least in the first stages of severe convection. To study the role of the deep modes on the initiation of the convection in this section, the results of the moist case experiments WETA and WETB, which employ the same two initial baroclinicities used in DRY3A and DRY3B, will be discussed. The character of the deep modes of these moist experiments will be compared with those of their dry counterparts in order to establish to what degree the organization of the convection is determined by dry atmospheric dynamics.

a. Experiment WETA

Clouds first appear at 1215 LST or 45 min after the three-dimensional simulation WETA was initiated from DRY3A. Before the deep modes are fully established, condensation occurs as scattered shallow clouds that cover the middle of the eastern slope, as viewed from the perspective of cloud-water content at 1300 LST (Fig. 14a). The clouds remain relatively shallow up to about 1400 LST during which time their areal coverage extends towards the western edge of the east-
ern slope. The gravity-wave and boundary-layer eddy structures are quite similar to those in the dry case during this period, indicating that the clouds are too small to significantly affect the overall dynamics. At 1400 LST, the first convective towers can be observed between 800–900 km (Fig. 14b). Deep convection, which moves eastward at the midtropospheric wind speed of 20 m s$^{-1}$, continues to the end of the simulation. Figures 14c and 14d show views of stronger convective activity at 1500 and 1530 LST. There is a rather abrupt eastern boundary to the region of moist convection that gradually moves eastward from $x = 950$ km at 1300 LST to $x = 1050$ km at 1530 LST. In DRY3A, the region east of $x = 1000$ km was one of the more active regions (Fig. 6d). These differences suggest that the influence of moisture sources as well as the presence of clouds has modified the dynamics. The following spectral analysis assesses this point in more detail.

Except for some minor differences at high wavenumbers, the power spectra of $w$ at 1310 LST for WETA (Fig. 15a) and DRY3A (Fig. 8b) are similar, indicating that the first small clouds have not significantly modified the gravity-wave structure. The situation is quite different once deep convection has developed. At 1440 LST, WETA shows a well-defined spectral peak at $\lambda_1 = 11.4$ km ($k_x = 35$) with about double the amplitude it attained in DRY3A. This differs considerably from the rather broad range (11 to 28.6 km) obtained in DRY3A. This selection is maintained until the end of simulation WETA, suggesting a well-established character to the convection during the last hour of the simulation. The peak at 28.6 km ($k_x = 14$) is about two-thirds of that found in the dry case. There is a significant energy peak at the $\lambda_1 = 6$ km ($k_x = 69$) in WETA that was not present in DRY3A. This difference is attributed to the presence of small shallow clouds in the western part of the eastern slope. It is worth noticing that, for experiment WETA, the maximum energy peak in $w$ has shown little change in scale between 1310 LST (10.8 km; $k_x = 37$) and 1440 LST (11.4 km; $k_x = 35$) in contrast to DRY3A where the larger-scale peaks at wavelengths of 28.6 km ($k_x = 14$) showed up during the later development of the deep modes. The spectral analysis of $q_c$ for WETA at 1440 LST (Fig. 15b) shows an even clearer spectral peak at 11.4 km than the analysis for $w$. Analysis of cross sections shows that the peak around 17.4 km ($k_x = 23$) is a result of the existence of anvils in the upper tropospheric levels.

In summary, "dry" deep modes are initially excited with horizontal wavelengths around 10–13 km and are found in both the dry and wet experiments. Then, if sufficient conditional instability is present, deep convection is triggered that appears to lock onto the "dry" deep modes existing at that time. Without clouds, the deep modal spectrum broadens, choosing larger horizontal scales ($\approx 28$ km). These results may have important relevance to the overall development of deep convection in nature. There are a number of factors that may delay the onset of cloud development in nature; e.g., the presence of inversions or dry layers. In such situations, this delay may allow for the development of dry deep modes with larger horizontal scales with the consequent development of deep clouds at these larger horizontal scales.

A more detailed analysis is presented between 1420 and 1445 LST over a reduced 100 km segment of the eastern slope (840–940 km) where the convective activity is intense (Fig. 16) in order to gain more insight into the nature of the moist convective and deep modal interactions in WETA. As previously discussed, deep clouds appear to lock onto preexisting dry deep modes.
The Rayleigh modes are not entirely suppressed over this region but the deep modes dominate in the vicinity of both clouds forming at $x = 900$ and 915 km. The development of intense convection is marked by the strongest vertical velocities at $z = 6$ km (Figs. 16d) with a maximum of about 14 m s$^{-1}$. Only the storm on the extreme right leads to surface precipitation. At this time, the clouds move at about the midtropospheric wind speed. Conversion of cloud water to rainwater occurs at about 1430 LST with precipitation reaching the ground at 1500 LST.

The behavior of the two clouds developing near $x = 890$ km (Fig. 16b) demonstrates how the presence of liquid water and the accompanying moist thermodynamics can modify the deep modal structure. The moist processes begin to strongly modify the updraft/downdraft structure from 1440 LST (Fig. 16d) until by 1445 LST the updraft of the eastern storm (near $x = 910$ km) is completely dissipated (Fig. 16e). These clouds then dissipate, leaving upper stratiform regions at the end of the simulation. This time evolution of $w$ also gives a more explicit picture of the modification of the original dry deep modes by moist convection. For example, the dry deep modes between $x = 890$
Fig. 14. Perspective view of surfaces of q_e > 0.01 g kg\(^{-1}\) on the eastern slope from experiment WETA. Times shown are (a) 1300 LST, (b) 1400 LST, (c) 1500 LST, and (d) 1530 LST.
and 920 km, present at 1420 and 1425 LST, are less pronounced at 1435 LST. This is perhaps a simple example of how clouds can modify the "dry" deep modes. Other more complicated ways clearly exist but have not been investigated.

b. Experiment WETB

Experiment WETB (the moist equivalent of DRY3B with a smaller initial baroclinicity of $3 \times 10^{-7}$ s$^{-1}$) shows an early behavior similar to WETA. Figure 17a shows a scattered shallow cloud field at 1300 LST that does not significantly affect the gravity-wave scale selection—as shown by the comparison of the power spectra of $w$ at 1310 LST between experiments DRY3B and WETB (Figs. 8d and 18a, respectively). However, the behavior of WETA and WETB at later times is quite different. As observed from the perspective of $q_c$ in WETB, the cloud field has a reduced horizontal extent and is much more confined to the west at later times. The main reason for this weaker cloud activity is the weaker amplitude of $w$ in the gravity waves in WETB than in WETA, as was also found for the dry experiments (section 4b). As in experiment WETA until 1400 LST, the cloud activity is confined to the western part of eastern slope. This is due to horizontal variations of conditional instability as well as variations of local convergence. These perspective views also show a broader range of selected horizontal scales of convective towers than in experiment WETA. As expected from the results of the dry experiments, the horizontal scale presently selected along the south–north direction is generally larger than in experiment WETA. The power spectra of $w$ and $q_c$ along the west–east direction (Figs. 18a and 18b) show that cloud and updrafts sizes are also typically larger than in WETA and extend over a larger range of scales. Dominant modes appear with $\lambda_c = 10$ to 16.7 km ($k_x$ from 40 to 24) with amplitudes about 1.5 times their dry-case values. In experiment WETA, this ratio was about 2. This relative difference in amplitude between the dry and moist experiments may be due to the reduced extent of cloud activity in WETB as compared to that in WETA. The magnitude of the 26.7 km ($k_x = 15$) peak is about the same in both the dry and wet experiments. This is in contrast to WETA where this large-scale secondary peak had a far smaller amplitude than in DRY3A. The later time scale selection of $w$ for WETB is much closer to the dry counterpart experiment DRY3B than was the case between WETA and DRY3A. The reduced baroclinicity in WETB results in weaker values of $w$ excited in the free troposphere and a slower rate of cloud development. The dry dynamics has a longer time to dominate in WETB than in WETA so that by the time the nonlinear cloud effects can influence the scale selection the deep modes are already well established. One might argue that the effective lock-on stage came much later in WETB than in WETA.

6. Summary and conclusions

Deep convection is often observed to develop in the absence of clearly defined forcing such as fronts or mountains. In such cases, local processes must initiate the deep convection. Thermally forced convection has been produced in both two- and three-dimensional numerical simulations using interactive grid nesting in a nonhydrostatic anelastic model. The outer domain of the model has an east–west dimension of 2400 km using 6 km horizontal resolution and with an inner domain of 1200 km using 2 km horizontal resolution to resolve the convective scale motions.

Convective initiation in our simulations was the result of free tropospheric gravity waves being excited by boundary-layer eddies penetrating the stable layer. One

![Figure 15](http://journals.ametsoc.org/journals/JAS/article-pdf/47/4/516/3425467/1520-0469(1990)047_0516_tiahss_2_0_co_2.pdf)

**FIG. 15.** One-dimensional power spectra for (a) $w$ and (b) $q_c$ over the eastern slope for experiment WETA at $t = 1310$ and 1400 LST. Here $k = 1$ refers to $\lambda_c = 400$ km.
mechanism believed to be particularly efficient is a Reynolds stress effect where the eddies act as small hills in the presence of mean wind shear. CHK, BC, and HC have studied such cloud excitation mechanisms but in much more confined domains. However, it is known that momentum mixing in the boundary layer tends to destroy the shear over a relatively short time scale and large-scale processes must be considered.
in order to maintain the shear against such mixing effects. A complete treatment of the local-initiation process requires the consideration of both small and large-scale effects in the same framework. Idealized initial conditions were used to gain insight into the processes of convection initiation. An idealized topography representing a 400 km long gentle western slope leading up to a 200 km wide plateau and followed by a 400 km long gentle eastern slope leading down to a flat plain, together with a constant west–east baroclinic
mean flow through a depth of 5.5 km, were used for the present simulations. These initial conditions together with a surface sensible heat flux lead to the creation of further shear in the region spanning the boundary layer top and overlying stable layer as a result of differential thermal gradients. More important, the physical processes creating such shear were found to be sufficient to maintain it against the effect of mixing by the boundary-layer eddies. The shear structure obtained in the first few kilometers of the atmosphere resulted in considerable horizontal variability over the domain. Stronger shear resulted in the first kilometers for the eastern slope region of the domain than occurred elsewhere.

Deep modes, which first developed along the eastern-slope region, selected west–east horizontal scales around 11–13 km in experiment DRY3A. Later in the simulation, a larger scale around 28.6 km was selected. Experiment DRY3B with a lower initial baroclinicity than in experiment DRY3A (3 × 10^{-3} vs 5 × 10^{-3} s^{-1}) showed a more banded structure. This difference was explained as a result of the spanning-shear structure being closer to pure directional as in BC. The west–east horizontal wavelengths selected were slightly larger than for experiment DRY3A. This difference was amplified in the two-dimensional simulations.

In experiment WETA, using the same initial conditions as experiment DRY3A but with the full warm-rain physics of the model, dry deep modes are initially excited with horizontal wavelengths around 10–13 km, similar to those found in the dry experiment. When deep convection is triggered it appears to lock onto the dry deep modes existing at the time of the first significant moist convection. As a result of this nonlinear scale selection, due to the presence of a field of clouds, the maximum horizontal scale selected was smaller in WETA (∼11.4 km) than found in DRY3A (∼28.6 km). The present results may have some important consequences for understanding the overall development of deep convection in nature. The horizontal scale sizes chosen by a cloud field may depend upon the stage in which the deep modes have developed before significant moist convection occurs. For example, weaker conditional instability may result in delayed onset of moist convection with the result of an increase in the horizontal scale. This appeared to be the case in experiment WETB, using the smaller initial baroclinicity. In this experiment, the clouds formed later in the simulation, due to weaker values of w excited in the troposphere. Larger horizontal scales were selected along both south–north and east–west directions than in experiment WETA. The dominant deep mode in WETB appeared with λ from 10 to 16.7 km as opposed to a 11.4 km for WETA.

Four main conclusions may be drawn from the results:

1) As found in previous studies, the magnitude and direction of the spanning-layer shear appear to be key parameters in selecting the spatial organization of cloud and vertical velocity fields. Such sensitivity was noted both within particular experiments as well as between experiments using different initial baroclinicities. The alignment of the boundary layer shear and overlying free-tropospheric shear is important in determining the three-dimensional character of the deep modes. Strong directional shear results in more banded (or two-dimensional) structures and unidirectional speed shear results in a scattered (or three-dimensional) structure. Such sensitivities to spanning-layer shear structure may play an important role in the atmosphere even when direct large-scale forcing mechanisms are present. This noteworthy sensitivity of modal response to the shear structure may allow for a characterization of the convection based upon the larger-scale flow structure. The classification of convective days during CCOPE by
LeMone (1989) supports this conclusion. Indeed, a strong correlation between the magnitude of the cloud base shear and the degree of convective organization was found in this recent study.

2) The time of onset of deep clouds relative to the stage of deep modal development may influence the horizontal-scale selection. In particular, delayed cloud formation may be followed by the development of clouds of larger horizontal scale, since the larger-scale dry gravity-wave modes will have had a chance to amplify.

3) The mean horizontal-momentum redistribution due to convectively forced gravity waves was found to be significant for the present experiments, even in the absence of clouds. Considering that large areas may have such convective wave activity, it may be important to consider such effects in the global momentum budget.

4) The simulation of the initiation of convection in undisturbed conditions as presented in this paper required sufficient horizontal resolution to resolve the dominant deep modes. The physics of this local initiation process was absent when 6 km horizontal-grid lengths were used but was "turned on" using 2 km grid sizes. This rather demanding resolution requirement has important consequences for the mesoscale modeling of air-mass convection.

In further extensions of the present work it may be instructive to explore the parameter space determined by varying the initial baroclinicity and conditional stability. Expanding the experimental framework to consider a more realistic topography and a larger domain in the south–north direction is also desirable.

As an extension of the present research, work is underway that concerns the explanation of the nocturnal convective activity maximum often observed on the Great Plains. Numerous studies have shown a strong correlation between the presence of the low-level jet and a convective activity maximum around midnight (Pitchford and London 1962; Bonner 1966; Maddox 1983) which is usually explained in terms of moisture advection. This correlation also suggests that the low-level jet may be a key ingredient in creating interactions between the boundary layer and overlying stable layer resulting in enhanced moist convection. Indeed, several such mechanisms have been proposed in the literature and will be explored. Gravity waves will be coupled with boundary layer eddies and excited in a manner similar to that described in this paper, except that the source of kinetic energy initiating the convection would originate in the free atmosphere. It should provide an alternate or complementary explanation of nocturnal convective activity in terms of forced dominant deep modes.

Acknowledgments. The authors wish to thank the referees and M. W. Moncrieff for their constructive comments. They also thank W. Hall for his valuable assistance with modeling problems. The present work was undertaken during J.L.R.'s long-term visit to the Mesoscale and Microscale Meteorology Division, NCAR, and was partly supported by the Institut National des Sciences de l'Univers, Paris, France under Grant PAM 89 36 06.

APPENDIX

Radiative Scheme

The source-sink term in the \( \theta \) equation due to the longwave radiative processes can be written as

\[
\left( \frac{\partial \theta}{\partial t} \right)_R = -\frac{\theta}{\rho C_p} \frac{\partial}{\partial z} (F_t - F_i),
\]

where \( F_t \) and \( F_i \) are the total upward and downward radiative fluxes.

Using the broadband emittance form, the upward and downward radiative fluxes at height \( z \) can be written respectively as (Stephens 1984):

\[
F_i(z) = \sigma T^4(z) [1 - \epsilon(z, 0)]
\]

\[
- \int_0^z \sigma T^4(z') \frac{\partial \epsilon(z, z')}{\partial z'} dz' \]

\[
F_t(z) = \int_z^\infty \sigma T^4(z') \frac{\partial \epsilon(z, z')}{\partial z'} dz'.
\]

In these expressions, \( \epsilon(z, z') \) is the integrated emmissivity along the optical path \( z - z' \). For homogeneous paths, the classical definition of \( \epsilon \) relates the emission of a column of atmosphere to that of a blackbody at the same temperature and is integrated through all the atmospheric wavelengths:

\[
\epsilon(z, z') = 1/\sigma T^4 \int_0^\infty A(z, z') \pi B_{\nu}(T(z')) dv,
\]

where \( A \) and \( B \) are respectively the absorptivity and the Planck function for wavenumber \( \nu \). It is generally difficult to obtain expressions of \( \epsilon(z, z') \) that are accurate enough to evaluate their derivatives with precision. To overcome this problem, the expressions (A-2) and (A-3) are generally integrated by parts to give

\[
F_i(z) = \sigma T^4 z_0 + \int_0^z \epsilon(z, z') \frac{d\sigma T^4(z')}{dz'} dz' \]

\[
F_t(z) = -\int_z^\infty \epsilon(z, z') \frac{d\sigma T^4(z')}{dz'} dz'.
\]

where

\[
\epsilon(z, z') = \int_0^\infty A(z, z') \pi \frac{dB_{\nu}(T(z'))}{d\sigma T^4(z')} dv.
\]
Since the shape of $dB_d/dT$ is similar to that of $B_r$, the quantities $\epsilon$ and $\epsilon'$ do not differ greatly (Stephens 1984). Equation (A-1) becomes

$$
\left( \frac{\partial \theta}{\partial t} \right)_{R} = -\frac{\theta}{T \rho C_p} \frac{\partial}{\partial z} \left[ \int_{0}^{\infty} \epsilon(z, z') \frac{d\sigma T^4(z')}{dz'} d\epsilon(z') \right] + \int_{z}^{\infty} \epsilon(z, z') \frac{d\sigma T^4(z')}{dz'} d\epsilon(z').
$$

(A.8)

In the troposphere, the principal effects of absorption and scattering are due to the rotation bands (18–100 $\mu$m) and vibration-rotation band (6.3 $\mu$m) of water vapor, the carbon dioxide band (around 15 $\mu$m), and the continuum absorption in the atmospheric window (8–14 $\mu$m). These effects together with the cloud-water effect are included in the present model, using the formulation proposed by Veye et al. (1980). In this formulation, $\epsilon(z, z')$ is expressed as a function of path length for water vapor and carbon dioxide, which are computed from integrals between $z$ and $z'$. The computation of double integrals [one to determine $\epsilon'$, a second to determine the fluxes through the expressions (A-5) and (A-6)] is obviously too expensive to use in a three-dimensional cloud model. A simpler and more economical formulation has to be found for practical purposes. The radiative effect at height $z$ may be decomposed into effects due to radiative exchanges between the height $z$ and (i) space, (ii) the ground, and (iii) the other layers. Assuming that the latter effect is smaller than the other ones, Sasamori (1972) has proposed to neglect this term. Assuming that the whole atmosphere has a temperature of height $z$, when evaluating the radiative flux divergence through (A-1), (A-5), and (A-6) and using Riemann–Stieltjes integration, the expression (A-8) becomes

$$
\left( \frac{\partial \theta}{\partial t} \right)_{R} = -\frac{\theta}{T \rho C_p} \left[ \sigma (T^4(z) - T^4_{z=0}) \frac{\partial}{\partial z} \epsilon(z, 0) - \sigma T^4(z) \frac{\partial}{\partial z} \epsilon(z, \infty) \right].
$$

(A.9)

In his boundary layer model Sommeria (1976), using

---

**Fig. A1.** Vertical cross-section of the radiative divergence at 0700 LST for simulation MESA with contour intervals of 0.5 K day$^{-1}$, computed from (a) the complete integral, (b) the new approximation, (c) the Sasamori approximation, and (d) the Sommeria approximation.
similar hypotheses but a slightly different derivation, approximated (A-8) by
\[
\left( \frac{\partial \theta}{\partial t} \right)_R^{SO} = -\frac{1}{T \rho C_p} \frac{\partial}{\partial z} \left[ \left( \sigma T^d(z) - \sigma T^d_{-\infty} \right) \epsilon(z,0) \right] - \sigma T^d(z) \epsilon(z,\infty). \tag{A.10}
\]

Both approaches fail to give a correct estimation of radiative divergence for the full troposphere. The approach used in the present work combines the full approach [Eq. (A-8)] and the approximated approach of Sasamori (1972) [Eq. (A-9)]. This new approach assumes that the radiative exchanges between the different layers of the atmosphere (term (iii)) cannot be neglected but the effect of their temporal variations may be neglected to determine the radiative divergence. The radiative term is written as
\[
\left( \frac{\partial \theta}{\partial t} \right)_R^N = \left( \frac{\partial \theta}{\partial t} \right)_R^{SA} + \left[ \left( \frac{\partial \theta}{\partial t} \right)_R^{SA} \left( \frac{\partial \theta}{\partial t} \right)_R^{SA*} \right]. \tag{A.11}
\]

On the right-hand side of this expression, the first term is computed at each time step by (A-9). The second term in brackets represents the difference between the complete form [Eq. (A-8)] and the approximate form [Eq. (A-9)]; i.e., the radiative exchanges between the different layers. This term is computed at regular intervals (in the present simulations, each 10 min). Figures A1(a)–(d) represent the vertical cross section of the radiative divergence at 0700 LST for the simulation MESA presented in section 3 for the approximations (A-8), (A-11), (A-9) and (A-10), respectively. The approximations of Sasamori and Sommer (Figs. A1c and A1d, respectively) roughly approach the complete formulation in the boundary layer, but largely fail in the upper layers. In contrast, the approach used in the present work (Fig. A1b) better approximates the full approach.

REFERENCES


