A Numerical Study of Inversion-Layer Breakup and the Effects of Topographic Shading in Idealized Valleys

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(Manuscript received 27 June 2002, in final form 7 March 2003)

ABSTRACT

Numerical simulations of inversion-layer breakup in idealized steep valleys are performed using the Advanced Regional Prediction System (ARPS) to investigate the effects of valley width and depth, and topographic shade. Simulations of the diurnal pattern of slope winds under weak synoptic conditions are presented in a valley of depth 500 m and floor width 1200 m. Typical up- and downslope wind circulations are reproduced, and their influence on the stability in the valley is analyzed and characterized using the classifications of Whiteman. A systematic investigation of the inversion-layer characteristics in a set of 24 valleys of varying depth and width is conducted. For the narrow-valley cases, the depth and lifetime of the stable layer increase as the depth of the valley increases. For wide valleys, however, the stable-layer depth and lifetime converge toward a single value regardless of the valley depth. An original subroutine accounting for topographic shading is introduced and its effects on both the slope winds and the inversion breakup process are discussed. Results from tests in idealized valleys indicate that topographic shading can delay inversion-layer breakup and, therefore, should be included, when appropriate, in numerical simulations of flow over complex terrain.

1. Introduction

A comprehensive understanding of the formation and breakup of stable layers in valleys is needed for prediction of the transport and mixing of pollutants over complex terrain (e.g., Whiteman and Barr 1986; Allwine et al. 1997; Savov et al. 2002). Over flat terrain, if synoptic conditions are weak, nighttime surface cooling leads to the formation of a stable layer, which inhibits vertical mixing. The stable layer is then destroyed in the morning by the growth of a convective boundary layer (CBL) induced by solar heating of the surface. In steep valleys, this diurnal pattern is strongly influenced by valley and slope winds.

At night, downslope flows affect the depth and strength of the stable layer. In the morning, upslope winds can prevent the convective boundary layer from growing but can also recirculate the warmed air over the slopes and induce a breakup of the inversion layer from above. To a lesser extent, along-valley circulation may also affect these dynamics. The effect of cross-valley circulation during breakup was investigated by Whiteman (1982), who classified inversion-layer breakup patterns in steep valleys. The first pattern is that observed over flat terrain (growth of a convective boundary layer). In the second pattern, often observed in very steep and narrow valleys, or when snow cover prevents surface heating, the convective boundary layer growth is limited and the inversion is actually destroyed by a secondary effect of the upslope winds: by removing the warmed air from the bottom of the valley, they lead to the sinking, and subsequent warming, of the inversion layer. The third pattern is a combination of the first two in which the inversion layer is destroyed from above and below. Whiteman (1982) observed this third pattern of breakup in 18 of the 21 cases that he documented, independent of the season, topography, synoptic conditions, or upper-level winds.

The influence of upslope winds on inversion-layer breakup has been extensively studied in field experiments [e.g., Whiteman 1982, 1990 (and references therein); Brehm and Freytag 1982; Müller and Whiteman 1988; Sakiyama 1990; Kuwagata and Kimura 1995]. A number of analytical and numerical models have also been used in the past to study inversion-layer breakup in valleys (McNider and Pielke 1981, 1984; Whiteman and McKee 1982; Tang and Peng 1983; Badger and McKee 1983, 1985; Kimura and Kuwagata 1995, Allwine et al. 1997; Anquetin et al. 1998; Li and Atkinson 1999). The aim of this paper is to use a large-eddy simulation numerical model to examine the influence of topography on the evolution of stable layers in several idealized geometries, from very steep to very...
wide valleys. Previous idealized numerical studies have investigated valleys of various shapes and sizes; however, the valleys are most often situated in the middle of a plateau region. In our simulations we have found the convergence of flow above the mountain ridges to be very important and have, therefore, chosen a topography that allows this.

In addition, a new topographic shading subroutine was developed, and the effect of topographic shading (which is neglected in most numerical models) is examined. Especially during early morning hours, topographic shade can play an important role in steep valleys in the development of slope winds, which then affect boundary layer evolution.

In the first part of this paper, we present a systematic numerical investigation of the influence of topography on the inversion-layer breakup in valleys of various aspect ratios. To achieve this, we use the Advanced Regional Prediction System (ARPS, version 5.0.0 Beta3c), a three-dimensional, compressible, and nonhydrostatic large-eddy simulation code developed by the Center for Analysis and Prediction of Storms, The University of Oklahoma (Xue et al. 2000, 2001). The ideal valley geometries are chosen such that convergence of upslope winds is possible along the mountain ridges, inducing recirculation into the valley from above. These simple valley geometries allow a study that isolates the influence of topography on the development and breakup of inversion layers. Observations show that during the early morning hours, cross-valley circulations prevail over along-valley winds (e.g., Kuwagata and Kimura 1995); therefore, the latter are neglected by using a flat valley floor. These simulations are not intended to be compared with any single set of observations, but, rather, should be viewed as a modeling tool for understanding the physics of inversion-layer breakup patterns in steep valleys.

In the second part of this paper, we focus on the effects of topographic shading. The inversion-layer breakup phenomenon with which we are concerned is primarily driven by solar heating of the ground in the morning. Whiteman et al. (1989a, 1989b) and Matzinger et al. (2003) emphasized the importance of the topographic shade, which, by delaying the local sunrise, strongly affects the net radiation balance. Whiteman (1982) observed a delay in local sunrise in steep valleys of up to 1 h 50 min after theoretical sunrise, depending on the orientation of the valley axis. The formation of a convective boundary layer usually followed approximately 0.5 h after local sunrise. These observations indicate that accurate computation of the incoming solar radiation in the early morning hours is critical for inversion-layer breakup simulations. Currently, ARPS only considers self shading (due to the orientation of a surface with respect to the sun) and does not take the topographic shade (blocking of solar radiation by neighboring topography) into account; to the knowledge of the authors, neither do any of the major comprehensive mesoscale codes. Therefore, we have added a new subroutine to the code to account for the topographic shade in the radiation balance. The method used is described, and test cases are analyzed in the context of idealized valley flows. The extent of topographic shading due to real topography from a Colorado valley is also demonstrated; however, simulations of a real case that could be used to compare with field data require accurate knowledge of synoptic and surface conditions and are left to future work.

2. Model setup

ARPS is extensively described in Xue et al. (2000, 2001) and will not be presented in detail here. The code has been previously validated using several symmetry tests and idealized geometries, in addition to being used for simulations with real topography and data (Xue et al. 2000, 2001). This section summarizes the configuration of important parameters used in our simulations of idealized valley flows.

a. Topography and grid setup

Most previous numerical studies consider a valley located in the middle of a plateau (e.g., Bader and McKee 1983; McNider and Pielke 1984; Anquetin et al. 1998). However, we found that a topography commonly used to study plain-to-basin flows (de Wekker et al. 1998) is needed to account for the convergence of upslope winds above the ridges. This convergence is influenced by asymmetric heating of the east and west slopes and leads to the destruction of stable layers from above [observed in patterns 2 and 3 of Whiteman (1982)]. Consequently, the basic topography consists of two triangular hills that form a valley of a trapezoidal cross section (see Fig. 1a). Several simulations have been performed with various widths of the horizontal valley floor $W$ and ridge heights $H$ while the half-width of the ridges remains constant at 2600 m. Before its use in ARPS, the terrain is processed using a four-pass Barnes scheme to smooth its sharp edges (Xue et al. 1995).

Kuwagata and Kimura (1995) and Whiteman (1990) observed that in the morning cross-valley circulation prevails over along-valley winds; Bader and McKee (1985) noted that along-valley winds did not affect the transition period of the boundary layer. These observations suggest that inversion breakup can be well-reproduced in a flat-bottomed valley. The floor of the valley is, therefore, chosen to be horizontal, allowing the use of a two-dimensional simulation. Simulations performed with a three-dimensional version of this topography (i.e., two parallel and uniform mountain ranges) produced identical results. Even in three-dimensional simulations, all fields remain uniform in the north–south direction throughout the simulation. The absence of along-valley winds in these three-dimen-
sional simulations is attributed to the symmetry and accuracy of the ARPS code; in the absence of along-valley forcing, as is the case here, no winds should be observed. Poulos et al. (2000) also found that a three-dimensional version of flow over a mountain ridge gave very similar results to their two-dimensional simulations. For the particular studies presented here, two-dimensional simulations are used.

The grid size (including boundary cells) ranges from \( nx \times nz = 67 \times 40 \) to 149 \( \times 40 \), depending on the width of the valley. The horizontal resolution is 200 m in all cases. The vertical grid spacing is 35 m near the wall and is then stretched using a hyperbolic tangent function. The average vertical grid spacing is 170 m, making the top of the domain approximately 6500 m above the valley floor. The top boundary condition was chosen to use linear hydrostatic radiation as well as Rayleigh damping, which was applied to the top 2000 m of the domain (Xue et al. 2000). Various configurations of the vertical grid, as well as different vertical domain extents and top boundary conditions were tested; however, no significant differences were observed in the simulation results. The east and west lateral boundary conditions are periodic. Together with the idealized topography used in the simulations, these numerical parameters create a two-dimensional flow in the east–west plane. The size of the valley floor at the east and west boundaries must be chosen carefully so that the topography is periodic in the east–west direction: the sum of the floor sizes on the edges must be equal to that in the center of the domain; otherwise, asymmetries in the valley geometries can drive a flow between the valleys, as observed by Hennemuth (1985).

Because several field studies investigating inversion breakup in valleys have been conducted in the Rocky Mountains (e.g., Whiteman 1982; Clements et al. 1989), Japan (e.g., Kuwagata and Kimura 1995), and the Alps (e.g., Hennemuth 1985), our theoretical topography is located at 40°N latitude; the longitude is 0°E so that local and UTC times are equal. For better comparison with elevated valleys, the valley bottom lies at 2000 m above sea level (ASL). All the simulations take place on 21 September, so that easier qualitative comparisons can be made with existing field observations, which are often performed in late summer.

b. Numerical parameters

The momentum and scalar advection terms are solved using a fourth-order horizontal and second-order vertical differencing scheme. The time-integration scheme is leapfrog with a 0.5-s time step for the advection term and 0.05 s for the acoustic modes. The time steps were estimated using the ARPS stability criteria (using the grid spacings given above and maximum values for the horizontal and vertical winds of 10 and 7 m s\(^{-1}\), respectively); these were then reduced to account for the influence of the topography, which makes the local grid spacing (and, hence, the required time step) smaller because of the use of sigma coordinates.

The subgrid-scale turbulence model chosen for the large-eddy simulation is the level-1.5 turbulent kinetic energy model. In ARPS this formulation has been altered for large-eddy simulation by requiring the length scale in the eddy viscosity relation to be connected to the grid spacing, so that the subgrid-scale model represents only the unresolved scales of motion [see Xue et al. (2000) for details]. We have also performed simulations using no turbulence model and using a stability-dependent version (nondynamic) of the Smagorinsky model. While there are small variations in the instantaneous fields, the overall evolution of the inversion-layer breakup is not greatly affected, so that the mean parameters, such as time for breakup and the depth of the inversion, are essentially unchanged.

The Coriolis force is neglected because of the small size of the domain. Moisture processes are activated, but microphysics are not included; warm (liquid) saturation adjustment is performed, and saturation and condensation processes are included. The activation of moisture processes allows a more realistic simulation, and is necessary for accurate soil moisture development, because even at high elevations moisture concentrations can affect the daily temperature cycle in the air and in the soil.

c. Initialization of wind and potential temperature fields

The model is initialized at 0500 UTC with conditions similar to those of Anquetin et al. (1998), with a slightly stable atmosphere imposed uniformly over the entire domain. The potential temperature near the ground is 296 K and the Brunt–Väisälä frequency is constant everywhere and is given by \( N = 0.019 \text{ s}^{-1} \) (corresponding to a potential gradient of 11 K km\(^{-1}\)).

Because calm winds are often observed in the morning in such valleys (e.g., Kuwagata and Kimura 1995), the model is initialized with zero winds. Tests with a slight westerly wind of 0.1 m s\(^{-1}\) added to the initial base state produced comparable results. McNider and Pielke (1984) and de Wekker et al. (1998) used a light initial synoptic wind, and Anquetin et al. (1998) used zero initial winds to isolate thermal effects in their simulations. Weak synoptic wind conditions are also of interest because they present a worst-case scenario for pollution problems.

The model was run for 48 h, but to minimize the effect of the initial conditions, only the second day of the simulation is analyzed. For the valley discussed in the following section (shown in Figs. 1a–g), the initially uniform and slightly stable (stability class 2; see Table 1) atmosphere evolves to become neutrally stratified during the first few hours. The neutral layer extends up to the first 1000 m above ground level (AGL) (stability class 3) and remains slightly stable above this level.
During the first night, a stable layer develops near the surface, extending about 600 m above the surface. Therefore, at the beginning of the second day, the potential temperature gradient is no longer neutral throughout, but consists of stable and neutral layers that are consistent with observations in steep valleys (Whiteman 1982). The resulting potential temperature profile after the initialization period (shown later in Fig. 3) is similar to that used by Bader and McKee (1983, 1985) to initialize their simulations.

**d. Surface characteristics**

For simulations of real cases, heterogeneous surface conditions would be used, because these will influence the development of the boundary layer. Particularly in valleys, surface properties at the valley floor will often be quite different than those high along the slopes. However, for our purposes, it is adequate to specify that these surface characteristics are uniform over the valley. Their values are chosen so that the modeled valley is similar to a typical Colorado valley. The surface characteristics of the Eagle, Yampa, and Brush Creek Valleys (Whiteman 1982; Clements et al. 1989) were obtained using the surface data preprocessor in ARPS. A rough average of these values was used to initialize our idealized valleys. The soil type is chosen to be loam, the vegetation cover is grassland with shrubs, the roughness length is 0.2 m, the leaf area index is 3, and the vegetation fraction is 0.5. See the description in Xue et al. (2001) for details on the surface characteristics database.

The behavior of the potential temperature fields in the atmospheric boundary layer is strongly influenced by soil moisture. The time-dependent soil temperature and moisture are predicted by surface energy and moisture budget equations, which include the effects of solar radiation, latent and sensible heat fluxes, condensation, and evaporation, among others (Xue et al. 2001). We have chosen the initial volumetric surface and deep soil moisture to be 0.13 m$^3$ m$^{-3}$, which is a typical observed value for this soil type [e.g., Noilhan and Planton 1989; similar values were also used by Xue et al. (2001) to simulate the Wangara experiment]. The deep soil moisture evolves very slowly over the simulation period, and its specific value does not have a strong impact on the evolution of the boundary layer over the timescales of these simulations. The surface soil moisture evolves during the initialization period of the simulation and stabilizes to a value of approximately 0.09 m$^3$ m$^{-3}$. The initial surface soil temperature is chosen to be the same as that of the first layer of air (initially 277 K), while the deep ground temperature is initially 15 K warmer (at approximately 292 K) to account for its role as a thermostat (i.e., in the early morning the deep soil will be much warmer than the surface). During the first 24 h of simulation, the deep soil temperature equilibrates to approximately 282 K, which lies in the middle of the surface soil temperature range of 267–300 K observed during the second day. The simulation results are especially sensitive to the choice of the surface soil moisture, because the observed range of surface temperatures becomes large if the surface soil is dry.

**3. Slope winds and inversion breakup in idealized valleys**

*a. Inversion breakup in an idealized valley*

We present here in detail a simulation of the diurnal circulation of slope winds and their influence on the stability of the atmosphere within a 500-m-deep and 1200-m-wide valley. This valley geometry exhibits an inversion-layer breakup that follows Whiteman’s pattern 3.

1) **WIND PATTERNS AND STABILITY CLASSES**

Figures 1a–g represent seven cross sections of the valley; the entire east–west computational grid is shown up to 4.5 km ASL. The contours of stability class, four potential temperature profiles, and the wind vectors (at every grid point) are plotted. The scale of the horizontal and vertical wind vectors differs and is given in the upper left-hand corner of each figure. The stability of the atmosphere is diagnosed from the vertical gradient of the potential temperature, using a classification based on that of Pasquill–Gifford (see Table 1). To give a better understanding of the evolution of the potential temperature, four profiles are plotted: at the top of the western ridge, along the east-facing slope, in the center of the valley floor, and along the west-facing slope. There is a variation of surface potential temperature with altitude so that the actual range of values varies with each profile; however, the scaling of each profile is the same and they are later collected together (see Fig. 3) to show the evolution of the potential temperature profiles over time at these four locations. In Figs. 1a–g the profiles are shown so as to illustrate their shape with relation to the stability; hence, the actual temperature values are not included here.

The center of the valley floor and both east-facing slopes are fully lit by the sun by 0550 UTC, initiating an upslope flow along the full length of both east-facing slopes. The slice in Fig. 1a is taken in the early morning.
(0800 UTC), approximately 2.5 h after theoretical sunrise. The stability has not yet been affected by the warming from the sun. The winds in the valley are very weak; upslope winds on the east-facing slopes reach a maximum of 0.3 m s\(^{-1}\) at the first grid level (30 m AGL).

By 0900 UTC, the solar heating of the ground leads to the growth of the upslope wind on the east-facing slope to 2 m s\(^{-1}\) near the ground. On the west-facing slope the upslope wind is much weaker (about 0.2 m s\(^{-1}\)) but already creates a convergence above the crests, inducing a small updraft approximately 200 m deep. Another consequence of the surface heating is the growth of a neutral layer (stability class 3) on the top of the east-facing slopes, which, together with the updraft of warm air on the ridges, destroys the shallow stable layer above the crest (see the potential temperature profiles). Apart from these local changes, most of the valley remains unperturbed at this time.

Figures 1c–d at 0930 and 1000 UTC clearly show that the stable-layer breakup follows Whiteman’s pattern 3 because it is destroyed from both the bottom (through surface heating) and the top (through the recirculation of warm air along the slopes into the valley from above, as well as the sinking and warming of the stable layer, induced by the removal of air from the bottom of the valley). Notice that the development of the CBL along the valley floor and slopes is clearly asymmetric; this can be compared with Fig. 15 of Anquetin et al. (1998) where such cross-valley variation is also observed. Kelly (1988) also observed asymmetric inversion-layer breakup in a wide Wyoming valley, hypothesizing the cause to be variation in surface heating dependent on the orientation of the slopes; our simulations confirm this.

By 1000 UTC (Fig. 1d), the upslope winds have developed on both sides of the valley, though they remain stronger on the east-facing slope (3.5 m s\(^{-1}\) vs 0.8 m s\(^{-1}\) along the west-facing slopes). Large circulating eddies (400 m deep) form above each ridge. Because the strength of the upslope winds is stronger on the east-facing slopes, these eddies present a slight asymmetry with respect to the crest. The neutral and unstable layers along the slopes increase in depth, while a shallow stable core remains in the valley. One-half hour later, the stable core disappears from the valley.

The fifth slice is taken at 1200 UTC (Fig. 1e); the atmosphere is neutral inside most of the valley, and an unstable layer is present adjacent to the slope. In the upper levels, the residual neutral layer grows and extends to 3500 m ASL. This growth is especially significant above the ridges where the convergence of upslope winds leads to a strong updraft extending up to 1000 m into the atmosphere at a maximum vertical velocity of 5.5 m s\(^{-1}\); the winds then turn back toward the valley when they reach the bottom of the upper stable layer.

By 1600 UTC (Fig. 1f), the atmosphere is homogeneously neutrally stratified over the first 1700 m, while the free troposphere remains stable above. The upslope winds continue to circulate along the slopes and above the ridges, creating large eddies of maximum velocity 6 m s\(^{-1}\). The well-organized structure of these eddies leads to a significant downdraft above the valley floor of vertical velocity \(-2.5\) m s\(^{-1}\). The circulation pattern of winds that we observe in Fig. 1f is very similar to the schematic diagram of Kuwagata and Kimura (1995, their Fig. 4), where upslope winds converge at the crests and return to the valley below.

Finally, in the evening at 2000 UTC (Fig. 1g), the wind becomes downslope at a velocity of about 2 m s\(^{-1}\) along the east-facing slopes and about 1 m s\(^{-1}\) along the lower part of the west-facing slopes, allowing a stable layer to return near the valley floor. The winds in the residual upper neutral layer are much calmer than during the afternoon. The downslope winds continue during the entire night, filling the valley with a stable layer, while the winds in the neutral layer calm down until the conditions return to the pattern of the first slice at 0800 UTC.

Our simulations show moderately strong vertical velocities due to the symmetry in the topography and the convergence of upslope winds above the crests; however, the slope winds compare relatively well with other simulations. In the simulations of Li and Atkinson (1999), the up- and downslope winds are lower (2–4 m s\(^{-1}\)) but are for much wider valleys with milder slopes. Using similar initial conditions and valley dimensions as in the present simulations, Anquetin et al. (1998) observe maximum wind speeds of about 7 m s\(^{-1}\). However, their valley geometry precludes convergence of vertical velocities at mountain peaks. In their study of plain-to-basin winds, where convergence can occur, de Wekker et al. (1998) observed 7 m s\(^{-1}\) upslope winds. Whiteman’s (1982) observations in several Colorado valleys showed upslope winds of up to 3 m s\(^{-1}\); the usual estimate for upslope winds is 1–5 m s\(^{-1}\) (Whiteman 2000, p. 186).

Over complex terrain, the presence of numerous mountain peaks would break the regularity of the convergence of upslope winds and, therefore, weaken the magnitude of vertical velocities in comparison with those observed in our idealized simulations. Furthermore, synoptic conditions would certainly affect the evolution of such large convective cells. For instance, while Bader and McKee (1985) noted that synoptic winds do not affect the transition period of the boundary layer very much, later in the morning these winds become important. In addition, Kuwagata and Kimura (1995) observed that in their field studies along-valley winds became important around noon.

The evolution over time of the stability class contours for this valley is also informative, because it clearly shows the pattern-3 breakup. Figure 2 shows a time series of the stability class versus height, and we can see the destruction of the stable layer from below by the growth of the CBL, as well as the destruction from
FIG. 1. The wind field (vectors), potential temperature (profiles), and stability class (filled isocontours; from 1: very stable to 6: strongly unstable) in an idealized valley ($H = 500$ m and $W = 1200$ m) at (a) 0800, (b) 0900, (c) 0930, (d) 1000, (e) 1200, (f) 1600, and (g) 2000 UTC. Wind vectors are scaled separately for each plot.
the top. Figures 1c and 1d show that the mechanism of destruction from the top is due to the combined effect of the sinking and warming of the stable layer, induced by the slope winds and their recirculation above the crests. The stability class for this figure is diagnosed from the potential temperature averaged above the valley floor in the east–west direction. This time series can be compared with Fig. 7 of Whiteman (1982) and will be discussed for other valley geometries later.

2) POTENTIAL TEMPERATURE AND TURBULENT KINETIC ENERGY DEVELOPMENT

The evolution of potential temperature over time at the four locations in Figs. 1a–g is also shown separately in Fig. 3 and can be viewed as a summary of the evolution of the valley boundary layer. These profiles can be compared with the potential temperature soundings given in Fig. 5 of Whiteman (1982) for pattern-3 break-

Fig. 2. Time series of the stability class averaged across the valley floor and plotted vs altitude in an idealized valley ($H = 500 \text{ m}$ and $W = 1200 \text{ m}$).

Fig. 3. Evolution of the vertical profiles of potential temperature in an idealized valley ($H = 500 \text{ m}$ and $W = 1200 \text{ m}$) at 0800 (bold), 0900 (dashed), 1000 (dashed-dot), 1200 (dotted), 1600 (bold dashed-dot), and 2000 (bold dashed) UTC. Locations for the profiles are (a) top of the western ridge, (b) middle of the east-facing slope, (c) bottom of the valley, and (d) middle of the west-facing slope.
up. Note that Whiteman’s soundings extend only to about 700 m above the ground surface and are for the morning hours from sunrise until the inversion-layer breakup. The profiles in Fig. 3 are given at the times at which cross sections were plotted in Figs. 1a–g and, therefore, begin at 0800 and end at 2000 UTC. If we examine the portion of Fig. 3c (at the valley floor) extending to about 600 m, we see the gradual warming of the surface until the valley atmosphere is fully mixed and the inversion has disappeared (about 1000 UTC). Later in the day, the atmosphere continues to warm with a fairly uniform potential temperature profile (1200 and 1600 UTC) extending about 1.5 km above the surface. Nighttime surface cooling becomes evident in the lower portion of the profile at 2000 UTC.

The potential temperature profiles for locations on the western ridge, and halfway up the east- and west-facing slopes, are given in Figs. 3a, 3b, and 3d, respectively. The pattern of evolution of the profiles is similar to that on the valley floor. Between 1000 and 1200 UTC, the atmosphere has been warmed considerably by the large convective cells that extend about 1.7 km above the valley floor. The range of potential temperatures observed is smaller for the profiles shown from the higher elevations (along the slopes and on the mountain ridge). The sharp gradients in the upper portions of the midday profile of Fig. 3a are due to the strong updrafts, which can be seen in Fig. 1e.

Figure 4 shows cross sections of potential temperature and turbulent kinetic energy contours during the early morning hours of 0800, 0900, and 1000 UTC during which time the inversion-layer breakup process occurs. At 0900 UTC, the surface heating induces deformations in the potential temperature contours, generating up-slope winds. This is similar to the deformation described in Fig. 7 of McNider and Pielke (1984). The up-slope winds are responsible for the larger values of turbulent kinetic energy seen along the east-facing slopes. The updraft recirculation has begun to mix out the potential temperature gradients at the peaks; the growth of these well-mixed regions is clear in the cross section at 1000 UTC. The mixed regions correspond well to stability class–3 areas in Figs. 1b–d.

b. Influence of the topography

We now investigate the effect of the valley topography by performing simulations of a set of idealized valleys of various depths and widths. The geometry of a valley has a strong impact on the evolution of the boundary layer; in these idealized simulations it is possible to isolate the effect of the topography and change it. The same initialization parameters (described above) are used to allow us to highlight the influence of the topography on the formation and destruction of the stable layer. The results of these simulations are summarized by comparing the depth, lifetime, and pattern of destruction of the stable layer.

1) Depth of the stable layer

The top of Fig. 5 represents the depth of the stable layer at 0700 UTC in 24 different valleys of various depths \( H = 200, 500, 1000, \) and 1500 m) and widths of the valley floor \( W = 400, 800, 1200, 2400, 4800, \) and 9600 m). The potential temperature field is averaged across the valley floor to obtain a single profile, from which the stable layer is then diagnosed as the layer of atmosphere having a vertical gradient of potential temperature higher than 5 K km\(^{-1}\) (stability class of 1 or 2). Results of identical simulations, taking into account the topographic shade, are also plotted and will be discussed later. In some cases the development of the stable layer did not allow a clear determination of its depth; in the three steepest cases \( H = 1500 \) m and \( W = 400, 800, \) and 1200 m), the stable layer is so deep that it fills the entire domain and no residual layer remains during the night. In these cases, the depth of the stable layer was approximated as the depth of the neutral layer that developed the previous day.

The depth of the stable layer shows an important sensitivity to the aspect ratio of the valleys. First, we observe that when the width of the valley is small, the influence of the topography is much stronger, so that the depth of the stable layer scales with the depth of the valley. Hence, in the steepest cases \( H = 1000 \) and 1500 m) the depth of the stable layer increases up to 900 and 1300 m, respectively, when the valley floor is very narrow. Second, when the depth of the valley decreases, the influence of the valley width is not as great. Thus, for the two shallowest sets of valleys \( H = 200 \) and 500 m), the depth of the stable layer is almost constant with respect to the width of the valley. Similarly, when the width of the valley increases, the depth of the stable layer shows less dependence on the surrounding topography and converges toward 500 m for all four depths of valley tested. For comparison, a single point corresponding to a simulation over flat terrain has been added to this plot. Here, the stable-layer depth is about 400 m, which is consistent with the value obtained in the most open valleys and with that over flat terrain [see, e.g., the Wangara experiment simulated by Xue et al. (2001)]. These results are also consistent with the observations of Whiteman (1982), who observed a range of ratios of the stable-layer depth to the valley depth of 0.53 to 1.10. It is common for shallower valleys to have stable layers that extend beyond the ridge tops, while in deeper valleys the inversion layer is often lower than the ridge tops, as observed here (Whiteman 2000, p. 175).

2) Lifetime of the stable layer

In the bottom of Fig. 5, we show the lifetime of the stable layer (in hours after sunrise at about 0540 UTC) as the time at which the gradient of the averaged po-
FIG. 4. The (left) potential temperature (K), and (right) turbulent kinetic energy (m$^2$ s$^{-2}$) and velocities (m s$^{-1}$) in an idealized valley ($H = 500$ m and $W = 1200$ m) at 0800, 0900, and 1000 UTC.
tential temperature has a slope of 5 K km$^{-1}$ or less (stability class 3 or higher) everywhere in the valley.

As one would expect, the behavior of the lifetime of the stable layer shows some common features with that of its depth. The stable-layer lifetime for all valley depths converges toward a single value (about 5 h) when the width of the valley increases. This value is consistent with the simulation over flat terrain and also with observations from the Wangara experiment (Xue et al. 2001). Whiteman (1982) found that the time after sunrise required for inversion breakup in 18 different Colorado valleys was usually between 3.5 and 5 h, with an average of 4 h 35 min. Sakiyama (1990) observed a breakup time of 3–4.5 h after theoretical sunrise for two
shallow valleys (depth of about 270 m and widths of 100 and 1200 m), and Müller and Whiteman (1988) found a breakup time of 4.5 h in a Swiss valley (depth of 1 km and width of 4-5 km at the crests). These observations compare quite well with the lifetimes of about 4.5-5.5 h for the 200-, 500-, and 1000-m-deep valleys shown in the bottom of Fig. 5. The steepest valley ($H = 1500$ m and $W = 400$ m) experiences a stable-layer lifetime of more than 6 h, which is comparable to the breakup time observed by Whiteman (1982) in the Eagle Valley.

Similar to the stable-layer depth, the lifetime of the stable layer increases when the depth of the valley increases. It is, again, in the narrowest case that this trend is the strongest, the breakup occurring about 2 h later in the 1500-m-deep valley than in the 200-m-deep case. For the shallowest valleys ($H = 200$ m), the lifetime shows a slight increase for all widths of the valley floor that were tested; in the deeper valleys ($H = 500, 1000$, and $1500$ m), the lifetime decreased (sometimes dramatically) as the valley width increased.

The two factors leading to the destruction of the stable layer are the growth of a convective boundary layer and the breakup of the inversion from above. In their simulations of valleys surrounded by plateaus, Bader and McKee (1985) found that the time required to break the inversion was longer for wider valleys of a fixed 500 m height. This trend is not observed here, except for the shallowest valleys ($H = 200$ m). The reasons for this may be in the choice of topography; the presence of plateaus can affect the recirculation of the slope winds back into the valley. (Tests with our code using this geometry confirmed the distinctive differences in the pattern of upslope wind flows observed by Bader and McKee; namely, winds traveled up the slopes and out along the plateaus for some distance before returning toward the valley.) The fact that Bader and McKee observe an increase of the lifetime of the stable layer when the valley widens indicates that the second factor (break up from above) may be more efficient in narrow valleys.

Our simulations confirm this; if we compare simulations in which the stable-layer depth is similar to that over flat terrain ($H = 200$ and 500 m), the time required until breakup is always smaller than when the growth of a CBL is the only driving factor, as in the flat case. However, our simulations show that the lifetime does not decrease systematically when the valley floor narrows; the observed trend depends on the depth of the valley, and the depth of the stable layer itself is likely a large factor in determining this trend. In the deepest cases, the recirculation of the slope winds takes longer to develop and may, therefore, act to increase the lifetime of the stable layer. The presence of plateaus affects the recirculation of upslope winds and may also be important in determining the trend with increasing valley width.

3) **Pattern of inversion-layer breakup**

Similar to Fig. 2, Fig. 6 represents time series of the stability class versus altitude in six different valleys ($H = 200, 500$, and $1000$ m; $W = 400$ and $9600$ m). The stability class is diagnosed from the potential temperature averaged above the valley floor (see Table 1). The data are sampled every 10 min during the simulation. This figure can be compared with Fig. 7 of Whiteman (1982). As described earlier for the valley in Figs. 1a-g, in each of these cases the initialization period is similar: a stable layer develops near the valley floor and is covered by a residual neutral layer, above which the atmosphere is not affected by the daytime warming of the first day and presents a stable stratification.

In the early morning hours, we sometimes observe very strong temperature inversions in the lowest layers near the surface, as did Whiteman (1982). The characteristics of this near-surface inversion layer (stability class 1) have not been discussed in detail in the previous sections because, as seen in Fig. 6, we do not observe the formation of this strongly stable layer in all cases. Indeed, it develops only in the more open valleys, and its depth decreases as the valley becomes more steep, until it disappears.

The depths and lifetimes of the stable layers discussed earlier can be seen in Fig. 6, where we emphasize the pattern of destruction of the stable layer, as classified by Whiteman (1982). In the very wide ($W = 9600$ m) or shallow ($H = 200$ m) valleys, the stable layer is destroyed primarily from the ground as it would be over a flat terrain (pattern 1). In contrast, in the narrow and deep valleys the breakup occurs mainly from the top. This shows that pattern 2 is also successfully reproduced, the deep stable layer being mostly destroyed by the secondary effect of the slope winds (the sinking and warming of air above the valley floor). This is particularly striking in one of the steepest cases ($H = 1000$ m, $W = 400$ m) where the convective boundary layer grows to 250 m while the top of the stable layer continues to sink, so that a stable core remains in the valley, significantly delaying the breakup. The descent of the inversion top is not as pronounced in wider valleys, which is consistent with the results of Bader and McKee (1985).

4. **Topographic shade**

a. **Implementation of the shading algorithm**

The current version of ARPS accounts for self shading in the radiation balance (i.e., the local surface slope is compared with the sun’s inclination angle), but not for the blocking of radiation induced by the topography (e.g., the shadow cast by a mountain). Whiteman et al. (1989a) emphasized the key role of direct radiation on the local circulation in elevated valleys (where the relative importance of diffusive radiation is weak) and showed how the direct solar beam is affected by the
topographic shading. The subsequent modification of the net all-wave radiation flux is presented in Whiteman et al. (1989b), where the predominant role of the direct solar beam on the overall radiation budget in a valley is identified. The recent field study of Matzinger et al. (2003) confirmed these observations and performed comparisons between overcast and clear days. During cloudy days, the incoming solar radiation is mostly diffuse and, thus, almost uniform in the valley. In contrast, the direct beam prevails on clear days, causing strong variations between the shaded and sunny locations. Whiteman (2000, pp. 181, 312–319) also discusses the effect of topographic shade on delaying sunrise in complex terrain and includes a subroutine to compute the delay in local sunrise induced by self shading.

We have, therefore, added a new subroutine to ARPS, which computes the topographically shaded areas in order to modify the radiation balance computation. The method consists of drawing a line between each surface node and the sun, whose position is known from its azimuth and declination each time the radiation balance is computed (in these simulations, every 5 min). If this line intersects neighboring topography between the surface node and the sun, the surface node is marked as shaded. The shading test is performed at all grid nodes. Assuming that the topographic shading will be used only for rather small domains, the azimuth and declination of the sun are considered to be constant over the simulation domain. Last, the radiation balance is modified so that, at a shaded point, the direct component of the incoming solar radiation is set to zero. This subroutine is now available in the latest version of ARPS. The new topographic shading subroutine does not add significantly to the computation time; the observed increase in computing time was less than 1%. Note that our use of the terms “nonshaded” or “without shading” implies the default situation in ARPS where the topography is self shaded. “Shaded” and “with shading” imply adding the topographic shade, which increases the asymmetry of the valley heating.

b. Influence of shading on the upslope winds

The simulations presented above (without topographic shade) have also been performed with the use of this topography-induced shading subroutine. For the two-dimensional ideal valley simulations, the shade computation was adjusted to behave as if the mountain ranges extended uniformly in the third direction. Note that for this idealized case, the code could also have been
modified to make the induced shading exactly periodic, for example, to artificially shade the easternmost east-facing slope in the morning. However, the intent in developing the topographic subroutine was for application to real topography simulations. In a real case using grid nesting, it will also be important to be aware that topography outside of the domain can cast a shadow into the domain, though it would be cumbersome to include this. The effects of ignoring shading near the boundaries would likely be negligible in comparison with other errors generated at the boundaries by grid nesting.

In steep valley cases, the topographic shade has a very strong influence on the onset of the upslope winds. We can examine the effect of shading for an idealized valley 1500 m deep and 800 m wide. First, the east-facing slope of the easternmost mountain begins to be illuminated at about 0540 UTC, initiating weak upslope winds along the face; this slope is lit immediately at theoretical sunrise, because there is no mountain to the east to shade it. Within 10 min, the peak of the westernmost mountain is illuminated. As time progresses, more of the east-facing slope on the westernmost mountain is lit by the sun. The region of upslope winds gradually increases from just near the peak to cover the entire slope. By 0720 UTC, both east-facing slopes are fully illuminated; however, the valley floor remains in the shade. The center of the valley floor does not experience sunrise until just before 0800 UTC, a delay of 2 h 20 min in comparison with theoretical sunrise. Whiteman observed a delay of up to 1 h 50 min in the local sunrise in his field experiments (1982).

This growth of the upslope wind region can be seen in Fig. 7 by examining the $-0.5$ m s$^{-1}$ contour of $u$ velocity at the first grid level along the westernmost east-facing slope versus time. Without topographic shading (top of Fig. 7), the upslope wind ($u < 0$) becomes faster than 0.5 m s$^{-1}$ at about 0745 UTC uniformly along the slope. When topographic shade is considered (bottom), the local sunrise time is a function of the altitude along the east-facing slope. Consequently, at the top of the ridge the wind becomes faster than 0.5 m s$^{-1}$ at about 0745 UTC uniformly along the slope. When topographic shade is considered (bottom), the local sunrise time is a function of the altitude along the east-facing slope. Consequently, at the top of the ridge the wind becomes faster than 0.5 m s$^{-1}$ at about 0745 UTC uniformly along the slope. When topographic shade is considered (bottom), the local sunrise time is a function of the altitude along the east-facing slope. Consequently, at the top of the ridge the wind becomes faster than 0.5 m s$^{-1}$ at about 0745 UTC uniformly along the slope.

### c. Influence of shading on the stable-layer breakup

Valley cross sections showing the stability class in one of the steepest valleys ($W = 800$ m, $H = 1500$ m)
at 1000 and 1040 UTC, with and without topographic shading, are given in Fig. 8. Here, the stability of the valley without shading (left) at 1000 UTC is very similar to that of the shaded case (right) at 1040 UTC. By comparing the potential temperature contours, we observe that taking into account the shade induces an important delay in the morning-time warming of the atmosphere within the valley. However, because these slices are taken more than 4 h after sunrise, there is no structural difference in the stable-layer breakup between the two cases, apart from the delay. The differences in the onset of upslope flow with and without shading described in the previous section do induce a structural difference between the shaded and nonshaded cases, but only in the very early morning hours; by 1020 UTC, the upslope winds on the east-facing slope are very similar in both cases (see Fig. 7). Moreover this large delay in the onset of upslope flow only concerns the first grid level; at higher altitudes the difference is not as strong as in Fig. 7.

The influence of topographic shading on the stable-layer breakup can also be seen in Fig. 5, where the characteristics of the stable layer for 24 different valley geometries with and without shading are summarized. In the steepest cases, the morning-time stable-layer breakup (bottom) is delayed by about 0.5 h when topographic shading is used. Because stable-layer buildup occurs at night, the stable-layer depth (top of Fig. 5) is not affected by topographic shading.

To summarize, the effect of topographic shading in our idealized valleys is important in the very early morning, when local sunrise is delayed up to 2 h 20 min in the steepest cases; the delay in the onset of upslope winds at the valley floor is approximately 1 h, and the breakup of the stable layer occurs up to 30 min later than in the nonshaded cases for the steepest valleys. Bader and McKee (1985) found no difference in stable-layer breakup using symmetric or asymmetric surface heating or using a different valley orientation for a 500-m-deep valley. This agrees with our observations in the shallowest valleys, but for steeper valleys the effect of topographic shade is significant. In addition, the 0.5-h difference we observe in these cases is between the self-shaded case (which is already asymmetric) and the topographically shaded case. As mentioned in the example with real topography in the following section, valley orientation will also affect the percentage of topographically shaded area and, hence, the extent of a delay observed in inversion-layer breakup. The difference between the delay in the breakup time (0.5 h) and that of the onset of upslope winds (1 h) is probably due to cross-valley mixing driven by asymmetric surface heat-
Fig. 9. Shading in Brush Creek Valley 30 min after local sunrise: self-shaded areas (black), topographically shaded (gray), and illuminated sites (white).
two-dimensional valleys using ARPS. Use of idealized terrain has allowed a detailed exploration of the influence of the topography and topographic shading.

As far as idealized cases can be compared with real data, these simulations show rather good agreement with field experiments (e.g., Whiteman 1982). The characteristics of depth and lifetime of the stable layer are reasonable. The three topographically influenced breakup patterns described by White et al. (1982) are successfully reproduced; their occurrence depends on the valley geometry. The valley geometries consist of two parallel mountain ranges, which allow convergence of upslope winds at the crests; the resulting updrafts generate a recirculation that affects the destruction of the valley stable layers. The growth of a convective boundary layer at the valley floor, the sinking and warming of the stable layer induced by the slope winds, and their recirculation above the crests are all together responsible for the breakup of the stable layer.

An analysis of the influence of the width and depth of 24 valley geometries has also been conducted. The characteristics of the stable layer show a strong dependence on the surrounding topography. In narrow valleys, the depth and lifetime of the stable layer increase with the height of the ridges. In very open valleys, the influence of the topography decreases and the depth and lifetime converge toward those of a stable layer over flat terrain. For a valley of fixed depth, the dependence on increasing valley width is not entirely clear. In the shallowest valleys, the lifetime of the stable layer increases with valley width; in the deeper cases the opposite trend is observed. The influence of the convergence and recirculation of slope winds on the stable-layer breakup may be one reason that the effect of valley aspect ratio is not uniform for all of the cases tested; however, the effect of valley geometry on stable-layer breakup deserves further investigation.

In addition to demonstrating the effects of topography on inversion-layer breakup, the purpose of this paper was to emphasize the importance of topographic shading, especially for fine-resolution small-scale simulations, which are becoming more commonplace as computer power increases. While it should intuitively be known that shading is more important in a narrow valley, this has never been quantitatively demonstrated, and topographic shading (blocking of radiation by neighboring topography) is not included in any of the major meso- or submesoscale codes (to the authors’ knowledge).

A new subroutine has been added to the ARPS code to account for the topographic obstruction of sunlight in the incoming radiation computation. The influence on slope flows in idealized valleys in the morning is investigated, taking into account that the topographic shade adds a significant offset to the starting time of the upslope flows in the bottom of the valley and, hence, delays the completion of the inversion breakup in the deeper valleys. The observed stable-layer depths are not changed by the topographic shading. The extent of topographic shading has been investigated in field studies (Whiteman et al. 1989a; Matzinger et al. 2003), which, together with our results, suggest that topographic shading should be included in numerical simulations of steep valleys, or more generally in simulations using fine resolution over complex topography, especially because the additional computation time is negligible. Shading due to real Colorado valley topography was computed, showing the topographic shading to cover approximately 22% of the domain near sunrise, in addition to the self-shaded areas. Simulation of a real case involves many external factors (e.g., external forcing must be applied in a nested grid situation, and surface characteristics vary in space and must be obtained from land use datasets) and is, therefore, left to future work.

Acknowledgments. We are grateful to Prof. Ming Xue and the ARPS team for their guidance and support. This work was supported by TotalFinaElf (A. Colette), the U.S. Department of Energy under the auspices of the Atmospheric Sciences Program of the Office of Biological and Environmental Research (R. L. Street), and by the National Science Foundation Grant ATM-0073395 (Physical Meteorology Program: R. R. Rogers, Program Director) (F. K. Chow and R. L. Street). Computation was done at Stanford’s McCuen Environmental Computing Center. The assistance of Megan Bela and Thibaut de Crisnay on this project is appreciated.

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