Daytime Heat Transfer Processes Related to Slope Flows and Turbulent Convection in an Idealized Mountain Valley

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(Manuscript received 29 December 2009, in final form 4 June 2010)

ABSTRACT

The mechanisms governing the daytime development of thermally driven circulations along the transverse axis of idealized two-dimensional valleys are investigated by means of large-eddy simulations. In particular, the impact of slope winds and turbulent convection on the heat transfer from the vicinity of the ground surface to the core of the valley atmosphere is examined. The interaction between top-down heating produced by compensating subsidence in the valley core and bottom-up heating due to turbulent convection is described. Finally, an evaluation of the depth of the atmospheric layer affected by the slope wind system is provided.

1. Introduction

Various thermally driven winds are known to occur in mountain valleys under clear sky and weak synoptic forcing. Slope winds, valley winds, and mountain–plain winds (Whiteman 1990, 2000) occur at different spatial scales, but they are all generated by pressure gradients produced by differential heating between adjacent areas. Like all breezes, they blow from regions of relatively cooler and denser air to regions of relatively warmer and lighter air.

At the smallest scale, slope winds respond to temperature differences between the air heated or cooled by a slope and undisturbed air at the same level (Prandtl 1952; Schumann 1990; Haiden 2003). At a larger scale, valley winds are generated by thermal imbalances between the core of the valley volume and the atmosphere above a nearby plain. A comprehensive description of the different phases composing the daily cycle of thermally driven flows in a mountain valley, with emphasis on the interaction between slope and valley winds and on the transition between phases, was given by Defant (1951).

In the present work, we concentrate on the daytime, and in particular on the processes that govern the heating of valleys and cause primarily upslope flows and, as a consequence, the upvalley breeze.

Indeed, a key role in building the thermal imbalances that lead to the occurrence of valley winds is played by the different heating processes to which the valley and plain atmospheres are subject during the day. In particular, it is commonly observed that the atmospheric surface temperature at the bottom of a valley increases more than above plains. The larger diurnal temperature increase is commonly explained by the volume effect theory (Wenger 1923; Steinacker 1984). The volume of atmosphere confined below a given elevated horizontal area is smaller in a valley because of the presence of reliefs. Consequently, assuming that the same amount of heat is provided during the day to the atmosphere in a valley and above a plain, a larger temperature rise will be observed within the valley, where the heated volume is smaller: hence the thermal imbalance.

The volume effect concept is based on an integral evaluation of the overall heating of the valley atmosphere. It suggests a plausible reason why valleys are characterized by a larger daily temperature range, but it does not allow us to understand in depth the details of the ongoing heat transfer processes.

In this respect, circulations related to slope winds are thought to play an important role. It has been highlighted that flux divergence at the valley bottom, generated by the onset of upslope flows, can be compensated by subsiding motions in the valley core (Whiteman and McKee 1982; Vergeiner and Dreiseitl 1987). Adiabatic compression related to subsidence is a significant contribution to the heating of the valley atmosphere, at least initially. Because air maintains its potential temperature θ during

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the descent, the vertical θ profile is displaced downward: therefore, in a stably stratified atmosphere, subsidence drives potentially warm air toward the valley (Rampanelli et al. 2004; Weigel et al. 2006).

It is also known that when upslope flows occur in valleys or in closed basins, the growth of the convective boundary layer (CBL) can be partially suppressed (Whiteman and McKee 1982; Kondo et al. 1989; Weigel and Rotach 2004). Similarly, the depth of the CBL can be subject to a localized depression at the foot of isolated slopes during the development of anabatic winds (De Wekker 2008; Serafin and Zardi 2010). Furthermore, not only is the mixing height lower in valleys or near slopes, but also the capping inversion layer above the CBL is often reported to be deeper and less stable there (Rampanelli et al. 2004). Finally, although θ is generally larger in the valley atmosphere than above a plain subject to equal thermal forcing, an elevated layer where the atmosphere is potentially cooler within the valley can sometimes be observed, as demonstrated through numerical simulations by Rampanelli et al. (2004).

To summarize, the typical daytime vertical profile of potential temperature within a valley is subject to several factors that can make it significantly different from above flat terrain (e.g., flow divergence at the bottom, high-altitude subsidence, CBL growth suppression, occasional high-level cooling). The purpose of this paper is to clarify and quantify the processes that affect the transfer of thermal energy in a mountain valley and contribute to the generation of its peculiar thermal structure.

Two large-eddy simulations (LES), with identical initial conditions and the same diurnal solar radiation cycle and surface energy budget model, are performed to simulate the daytime flow field in idealized valleys with different cross sections. In particular, a narrow and a wide valley cross section are considered, with dimensions representative of typical Alpine valleys.

The paper is structured as follows: section 2 contains details about the numerical model in use, the setup of simulations, and the treatment of data. Section 3 describes several features of the flow field in valleys and explains their impact on the atmospheric heating process; dynamics occurring in the narrow and wide valley simulations are treated in parallel. A quantitative discussion is provided in section 4, aided by a simple Lagrangian analysis; the properties of selected air parcels are tracked along a few representative trajectories. Conclusions are drawn in section 5.

2. Numerical model

The Advanced Regional Prediction System (ARPS) model (Xue et al. 2000, 2001), already successfully used

in an LES setup for the simulation of thermally driven flows in real valleys by Chow et al. (2006) and Weigel et al. (2006), has been adopted for the present study.

a. Simulation domains

LES of the development of the CBL and of upslope flows were performed in two domains representing 2000-m-deep valleys with 30° slopes. The first valley has a triangular and the second a trapezoidal cross section (see Fig. 3): the two slopes are connected at the bottom in the first valley, while they are separated by a 5000-m-wide horizontal floor in the second. The slopes and floor are connected by curvilinear segments, each with a length approximately one-tenth of that of the slopes. Hence, the width of the two domains is of 8400 and 13 600 m respectively. The depth and width scales of topography in the two domains are roughly representative of the typical size and shape of Alpine valleys, respectively in their upper V-shaped part and in their lowermost regions, where the deposition of river sediments typically creates a flat alluvial floor between the sidewalls.

Simulation domains are three-dimensional: their extension in the along-valley direction is of 5000 m, so as to capture the dynamics of the largest convective eddies. For convenience, in the remainder of the paper the longitudinal valley axis is referred to as the x axis, and the cross-valley axis as the y axis.

The depth of the domains is of 8000 m, adequate to accurately represent the thermal plumes that develop at the top of valley walls and their overshooting into the free atmosphere up to a height of about 4000 m.

Note that the simple constant-cross section valley geometry outlined above does not allow us to appreciate the thermodynamic effects of the onset and propagation of the valley breeze, one of the major features of the flow field in valleys. However, it is known that in a valley with horizontal floor the upvalley breeze originates at the valley mouth before propagating both downstream and upstream as the diurnal heating cycle progresses (Rampanelli et al. 2004; Schmidli and Rotunno 2010). Accordingly, at any point sufficiently far from the opening of a valley onto a plain, a relevant fraction of the daily phase will be characterized by a totally absent or negligible valley breeze. Under this perspective, the results we describe below are representative of what happens in the interior of mountain ranges during the time lag between the early morning and midafternoon, when the valley–plain thermal contrast builds up as an effect of slope flows.

b. Atmospheric model

A fully compressible primitive equation system is used in ARPS: it includes prognostic equations for the three

components of momentum, for potential temperature and for pressure. The model setup is analogous to previous simulations by Serafin and Zardi (2010). Coriolis effects were not considered given the small spatial scale of the simulation domains. Subgrid-scale (SGS) diffusion terms were parameterized with a 1.5-order closure approach (Moeng 1984); hence, a prognostic equation for the SGS turbulent kinetic energy (TKE) was also integrated.

Since we aim at understanding only the basic features of the heating process, we consider a dry atmosphere in order to remove from our analysis any effect related to latent heat release, cloud cover, and deep convection. Therefore, no prognostic equations for water substances are used.

The ARPS model adopts a stretched terrain-following coordinate system: the minimum vertical grid spacing is 20 m at the surface and increases along the vertical up to about 125 m at the model top. The horizontal resolution is 50 m in both directions. The ratio between the horizontal and vertical grid spacing is such that the risk of an inaccurate computation of pressure gradients near sloping surfaces is minimized (De Wekker 2002). Arakawa C-grid staggering is adopted, and a third-order accurate advection scheme is used for momentum (fifth-order accurate for scalars). An explicit time-split leapfrog scheme with a Robert–Asselin filter is used for time integration except for the acoustic terms in the w and p equations, which are treated using the Crank–Nicolson implicit scheme. Weak divergence damping is implemented in order to minimize the generation of physically insignificant fast waves. Fourth-order computational mixing is also used to diffuse numerical errors and preserve computational stability.

c. Initial and boundary conditions

The initial condition of all simulations is a linearly stratified atmosphere at rest ($\partial\theta/\partial z = 3 \times 10^{-3} \text{ K m}^{-1}$). For simplicity, all possible modifications of this basic state (e.g., a nocturnal ground-based inversion or a residual mixed layer within the valley core) have not been considered for the purpose of the present study. A random potential temperature perturbation with a maximum amplitude of $2 \times 10^{-2} \text{ K}$ has instead been added to the initial field in order to trigger the startup of turbulent convection.

Boundary conditions are periodic on all lateral boundaries. Because of the approximate symmetry of the flow field along the y axis, the effect of periodicity is analogous to that of rigid wall conditions: indeed, very little horizontal motion across the domain boundaries at mountain tops was observed throughout the simulations.

Both the domain top and ground surface are treated as impermeable boundaries. A 2000-m-deep Rayleigh

damping layer is included below the domain top (8000 m) to absorb vertically propagating internal waves and prevent their spurious reflection.

Both simulations consider an idealized diurnal cycle of solar radiation. For convenience, the x (along-valley) axis has been set along the east–west direction, the domain center at a 0° latitude, and the simulation date in coincidence with an equinox. With these settings, the modeled solar path lies on a vertical plane parallel to the along-valley axis and produces symmetric radiative forcing over the valley walls.

We chose 0900 LST as the initial time of the simulations, in approximate coincidence with the moment at which the sensible heat flux first becomes positive in the diurnal cycle. This late initiation of the diurnal heating phase depends on the dry, desertlike conditions we are considering: in this case, most of the available net radiation feeds the ground heat flux rather than the sensible or latent heat fluxes, at least initially (Oke 1987). Only after several hours of radiative forcing does the ground surface become hot enough to initiate turbulent convection by buoyancy effects, thereby enhancing the heat transfer toward the atmosphere. Another relevant factor is related to the absence of an ambient wind in our analysis. Shear turbulence production is ineffective under calm winds, causing the heat flux from the ground surface to build up only after buoyant turbulence production has started: a condition that, as previously explained, occurs only in a few hours after sunrise.

d. Surface energy budget model

Since surface energy budgets play an important role in the process under investigation, the simplified model in use is described here in some detail.

Radiative forcing within the atmosphere is accounted for by semiempirical functions in the surface budget. The values of the parameters of the model are set so as to obtain reasonable diurnal cycles of the energy budget terms, even neglecting moist processes. The solar radiation reaching the earth surface is computed on the basis of the solar constant, by evaluating the effect of a variable zenith angle and a transmissivity function that accounts for Rayleigh scattering and absorption by permanent gases. No absorption by water vapor or cloud layers is considered. The slope angle and orientation are also accounted for in computing the irradiance of the solar radiation at the surface.

The net global radiation reaching the surface results from a balance between the absorbed solar radiation and the absorbed and emitted longwave radiation. The surface albedo and the emissivity of clear air and ground are set at constant values (0.725 and 0.995, respectively).

Land surface processes governing the energy exchange between soil and atmosphere are then treated using a simple two-layer force-restore model (Ren and Xue 2004; Noilhan and Planton 1989). For dry and bare ground the model only includes equations for the soil temperature, and the latent heat flux does not appear among the forcing terms at the surface.

The vertical component of the surface sensible heat flux H_0 is modeled using a bulk relationship:

$$H_0 = \rho_a c_p C_d U_a (T_s - T_a), \quad (1)$$

where $U_a = \sqrt{u_a^2 + v_a^2}$, u_a , and v_a are the horizontal wind components at the lowermost model level, c_p is the specific heat of air at constant pressure, ρ_a and T_a represent respectively the near-surface air density and temperature, and T_s is the soil temperature. The drag coefficient C_d is stability dependent (Xue et al. 2001). A similar parameterization is used for the surface momentum flux.

This formulation of surface fluxes is perhaps the main weakness of the model setup we chose. First, similarity functions included in the stability dependent formulation of C_d were actually devised in the Monin–Obukhov theory framework. Evidences for a possible extension of this theory to cover surface layer processes on the flat floor of wide valleys were shown by de Franceschi et al. (2009). However, its applicability to narrow valleys or the sidewall slopes is still questionable.

Second, bulk formulations of surface fluxes such as Eq. (1) can cause incorrectly low forcing in regions where the mean wind at the lowermost model level U_a approaches zero. This problem can be avoided in model runs with fully parameterized turbulence by adding a parameterized gustiness wind component to U_a (see, e.g., Beljaars 1994; Serafin and Zardi 2010). This solution does not seem to be appropriate in LES, where surface wind gusts are explicitly resolved. Near-zero surface forcing occurs in LES runs only in the short time lag before the onset of turbulent convection, without affecting the subsequent evolution of the flow field. In any case, we take simulation results with caution and try to recognize and discard any spurious effect related to model shortcomings.

e. LES averaging

The horizontal and vertical resolution of the simulations are fine enough to capture the dynamics of the largest boundary layer eddies [similar investigations of the PBL properties in mountain valleys—e.g., the LES by Chow et al. (2006)—adopted horizontal resolutions on the order of 150 m]. As is customary in interpreting the results of LES, the analysis is performed not on

instantaneous fields but on suitable averages in space and in time. The averaging operation aims at providing an estimate of the ensemble mean of turbulent quantities and is usually carried out along multiple dimensions to improve its significance (e.g., along x , y , and t to obtain a vertical profile of the quantity of interest). Accordingly, we computed averages along the longitudinal valley axis (x) and in 10-min periods, short enough to consider the flow field quasi-steady (the original data output frequency was 30 s).

All cross sections (Figs. 2–5), profiles (Fig. 7), and time series (Figs. 1 and 8) presented in this paper descend from averaging along the x axis and in 10-min periods, except where explicitly stated (Fig. 6). Where necessary, in order to obtain a vertical profile representative of the whole valley volume (Fig. 10), averaging along the cross-valley direction was performed horizontally, after transforming the model results from a terrain-following to a Cartesian reference frame.

3. Daytime heating of the atmosphere in a valley

In this section we first consider the surface forcing that generates anabatic motion and turbulent convection, and then proceed to a description of several localized features of the flow. Finally, we clarify the impact of the circulation on the thermal structure of the valley.

a. Surface fluxes

The time variation of the surface boundary conditions affecting the development of the boundary layer in the two valley runs is depicted in Fig. 1. The cycle of H_0 displays a delay and an attenuation in amplitude, compared to the solar and net radiation cycles. As already mentioned in section 2c, the phase delay can be explained considering the peculiar thermal properties of the dry and bare ground adopted in our simulations, in connection with the absence of an ambient wind. Solar and net radiation peak respectively at 1100 and 600 W m^{-2} around 1200 LST at the ridge tops and at the valley floor, while H_0 reaches a maximum of 500 W m^{-2} around 1400 LST.

Cross-valley inhomogeneities are apparent in the surface energy budget in the valley runs. The incoming solar radiation and, consequently, the net radiation are larger in the flat areas at the valley bottom and ridge tops, which are exactly perpendicular to the plane defined by the sunlight beams. Note that H_0 is also nonhomogeneous because of its dependence on the near-surface wind field. In the narrow valley it is larger above the slopes, where the upslope wind blows almost constantly, and smaller at the valley bottom, where the mean wind speed approaches zero for symmetry. Similar contrasts were also found in field measurements (see, e.g., Rotach and Zardi

FIG. 1. Time series (12 h, x axis) of relevant quantities of the surface energy budgets across the width (y axis) of the two valley domains, (top) narrow and (bottom) wide. Isolines are every 100 W m^{-2} for energy (zero contour marked with a thick black line) and every 0.1 m s^{-1} for momentum flux.

2007, in particular their Fig. 1a). Inhomogeneities are less pronounced in the wider valley.

The surface momentum flux, quantified by u_* in Fig. 1, is generally larger above the slopes (0.5 and 0.4 m s^{-1} in the narrow and wide valley respectively) than at the valley bottom (0.2 m s^{-1} in both valleys). The order of magnitude of modeled values is in good agreement with field measurements in real mountainous areas, both at the valley bottom (e.g., de Franceschi et al. 2009) and on the sidewalls (Andretta et al. 2001).

b. Flow features

The wind field in the valley cross sections is characterized by a series of localized features, including 1) upslope flow in the vicinity of the slopes, 2) thermal plumes detaching from the sidewalls at mountain tops, 3) horizontal cross-valley flow at ridge-top level, 4) subsiding air

motions in the middle of the valley, and 5) a CBL growing above the valley floor (Figs. 3–6; see also the sketch in Fig. 9, discussed in section 4a).

Figure 2 shows the slope wind field at an altitude of 1500 m , which is roughly representative of the upper two-thirds of the slopes: the main features of the upslope flow are different both from the steady-state Prandtl (1952) model and from Schumann's (1990) simulations. First, no downslope compensation current is clearly visible. Then, despite the impermeability condition enforced at the ground surface, a weak slope-normal wind component can be observed, directed toward the slope close to the ground ($\approx -0.1 \text{ m s}^{-1}$) and away from it at larger distances ($\approx 0.2 \text{ m s}^{-1}$). The latter is probably related to the slope-normal component of midvalley subsidence. The maximum upslope wind intensity is approximately 5 m s^{-1} (at 1400 LST) and 2.4 m s^{-1}

FIG. 2. Time–distance series of the along-valley (x -) averaged upslope wind speed (parallel and normal components) on the southern slope of the (top) narrow and (bottom) wide valleys. Surface altitude is 1500 m; isolines are every 0.4 and 0.1 m s^{-1} for the slope-parallel and slope-normal components, respectively. Gray lines denote negative values. Distance from the slope is evaluated along the slope-normal direction.

(at 1100 LST) in the narrow and wide valleys, respectively. The rather weak upslope flow intensity observed in the wide valley matches results from an LES of the evolution of the anabatic wind system above a 30° isolated slope presented by Serafin and Zardi (2010).

Although the two valley walls have the same slope angle and the initial atmospheric stability is the same in the two cases, different environmental conditions cause the upslope flow to evolve differently in the two model runs. In particular the atmosphere in the wide valley is subject to stronger mixing, which mitigates the thermal contrast between the heated slope and the valley atmosphere, and consequently weakens the upslope flow. This will be shown in more detail in section 4a.

In both the narrow and wide valley, the upslope flow layer deepens and decelerates in time, as turbulent mixing increases in the valley atmosphere. Also, an initially constant upslope flow is progressively replaced by a less regular flow field: after some time, both the slope-parallel and slope-normal wind components start being subject to rapidly alternating accelerations and decelerations. This indicates the transition from a quasi-steady anabatic flow

to an intermittent regime consisting of a succession of convective eddies running upslope. The transition is particularly evident in the narrow valley, where it occurs around 5 h after the beginning of the simulation (i.e., around 1400 LST). A smoother transition occurs around 1200 LST in the wider valley.

The mean wind field in the two valley cross sections is shown every 4 h in Fig. 3 (horizontal wind component v) and Fig. 4 (vertical wind component w). Because of the symmetry of topography and surface forcing, the flow field in both valleys is initially symmetric. Symmetry is substantially maintained on the slopes throughout the simulations but not above the floor in the wider valley, where randomly spaced thermals develop.

Upslope flows detach from mountain tops and create thermal plumes, with a maximum updraft speed of about 3 m s^{-1} and downward motions of about $1\text{--}1.5 \text{ m s}^{-1}$ at their sides in the narrow valley. The intensity of plumes is about half as much in the wide valley.

While the upward motion in thermal plumes is purely vertical close to the ridge top, the compensating flow has a substantial horizontal component. Even away from the

FIG. 3. Consecutive longitudinal mean fields of the cross-valley (v) horizontal wind speed component in the (left) narrow and (right) wide valley runs. Isolines are shown every 0.2 m s^{-1} ; positive and negative contours are black and gray, respectively.

plume, the wind blows horizontally toward the center of the valley, with an intensity of approximately 1 m s^{-1} . Therefore, warm air fed into the plume by the upslope flow detains laterally, and its thermal energy is advected toward the valley center in a relatively shallow layer that is slightly above the mountain-top level.

Mass continuity requires that the upward motion associated with the upslope flows be compensated by descending flow elsewhere. Figure 4 shows that subsidence indeed occurs in the valley atmosphere, at altitudes below the top of the thermal plumes over the mountain tops. Although subsidence initially affects only the lower portions of the valley atmosphere, at later times it extends up to (and beyond) the mountain-top level. It is stronger in the narrow valley (although the downward vertical wind speed never exceeds 0.3 m s^{-1}), and less pronounced ($<0.1 \text{ m s}^{-1}$) in the wider one, where weaker thermal

plumes are separated by a larger distance, thus requiring less intense compensating flow.

Figure 4 also shows that a CBL develops at the floor of the wider valley. Unevenly separated ascending and descending thermals can be observed, reaching a maximum vertical wind speed of about 1 m s^{-1} and a vertical extent between 1000 m (at 1300 LST) and 1500 m (at 1700 LST). In contrast, the flow field close to the narrow valley floor is dominated by flow divergence at the surface, balanced by downward motion higher up.

The cross sections of the resolved TKE field displayed in Fig. 5 confirm that turbulence is better developed above the floor of the wider valley, where the vertical extent of the CBL grows in time up to approximately 1500 m at 1700 LST. TKE peaks at about $2 \text{ m}^2 \text{ s}^{-2}$ at 1300 LST. Interestingly, the depth of the CBL appears to be slightly smaller above the bases of valley walls: this

FIG. 4. As in Fig. 3, but for w .

is an example of the occasional CBL depth depression that is likely to occur in the vicinity of slopes, according to De Wekker (2008). Turbulence is weaker ($\text{TKE} < 0.6 \text{ m}^2 \text{ s}^{-2}$) above the floor of the narrow valley: in that case, the valley core remains almost entirely nonturbulent.

The suppression of turbulence above the floor of the narrow valley can be explained considering advection effects. Midvalley subsidence brings nonturbulent stable air downward ($w < 0$, $\partial \text{TKE} / \partial z < 0$); meanwhile, persistent flow divergence at the valley bottom continually removes the TKE that is produced locally by buoyancy. However, it has to be noticed that a small underestimation in the buoyant production of turbulence, related to the surface sensible heat flux formulation, might also be caused by the parameterization in Eq. (1).

Common to the two valleys is the efficient buoyant generation of turbulence within the thermal plumes at the mountain tops. In both cases, TKE is not confined to the updraft region but is transported horizontally toward

the valley center by the mean flow. In the narrow valley, this mechanism creates a distinct elevated turbulent layer at an altitude slightly higher than the mountain tops. Turbulence is not generated locally there, either by buoyancy or by shear production: it is merely advected by circulations produced by the upslope flows.

The roles of vertical and horizontal advection of turbulent air after its detrainment from thermal plumes are very similar to the processes observed by Serafin and Zardi (2010) in their idealized simulations of upslope flows in a similar topographical setting. A detailed evaluation of the turbulence and heat budgets is also provided therein.

c. Heat transfer processes

The potential temperature field is modulated by all of the processes described above. Instantaneous θ cross sections, taken along a plane normal to the valley axis and

FIG. 5. As in Fig. 3, but for the explicitly resolved TKE. Isolines are every $0.2 \text{ m}^2 \text{ s}^{-2}$.

located at the middle of the simulation domain, are shown in Fig. 6. Subgrid-scale TKE is represented with a gray shading in the figure in order to clarify the role of turbulence in mixing the thermal profiles in different regions of the valley atmosphere.

Heating is initially most effective at the valley bottom and at crest height, where the spacing between isentropes increases, as is visible at 1300 LST. This is an indirect effect of the onset of anabatic winds. Stationary plumes, created at the mountain tops when air warmed along the slopes detaches from the ground, act as local elevated heating sources. They behave exactly as the thermals that cause entraining convection at the CBL top above a flat region: air parcels embedded in thermal plumes rise up to and beyond their level of neutral buoyancy (LNB); they are subject to upward buoyancy force until the LNB, but they overshoot because of their inertia; above the LNB they get mixed with potentially warm air from the free atmosphere above and undergo

a further small warming. From then on, their motion can be considered as adiabatic (see also section 4c).

Because of the adiabatic nature of the motion, the compensation of updrafts is parallel to isentropes, which are still unperturbed in the valley core and therefore still horizontal.

The net result of this process is that the heat input provided along the slopes actually enters the valley atmosphere at a height slightly above the mountain tops. This feature recalls the depiction of the return branches of slope circulations in the widely cited diagram by Defant (1951). However, in the present simulations horizontal compensation does not occur below the mountain tops, as suggested by Defant (1951), but rather in a layer located between the mountain tops and the LNB of plumes.

Based on our understanding of the physics behind this phenomenon, we remark that compensation is expected to be exactly horizontal only if the thermal updrafts above

FIG. 6. Consecutive middomain cross sections of θ (isolines every 0.2 K, thick every 1 K; $\theta = 312$ K at top) and subgrid-scale TKE (shading) in the (left) narrow and (right) wide valley runs. Subgrid-scale TKE, approximately one order of magnitude smaller than the explicitly resolved fraction, is used merely to suggest the position of the dominant turbulent eddies.

the opposite valley walls are of comparable strength (i.e., only if the valley walls are approximately symmetric in shape and altitude and warm the atmosphere with comparable heat fluxes, as in the present simulations).

Beyond heat, cross-valley flow from the mountain-top plumes also transports TKE along the horizontal. Advected turbulence is quite weak, but it is sufficient to make the thermal profile at crest height progressively less stable, as testified by the vertical spacing between isentropes increasing in time, as shown in Fig. 6.

Below crest height and above the CBL over the valley bottom, air subsides to compensate the upslope flow. The valley core is not affected by turbulent convection and therefore is not diabatically heated. Rather, air sinks adiabatically there, and the stable θ profile is displaced downward in analogy to the suggestions of Whiteman

(1982). This can be appreciated by looking at the 308.2–308.6-K isolines in the panels on the right of Fig. 6: while they initially lie around 2000–2500 m above the valley floor, they reach an altitude between 1000 and 1800 m at the end of simulations.

Figure 4 suggests that sinking motion occurs rather uniformly across the valley width. However, mass conservation requires air parcels to accelerate while getting to lower heights, because the horizontal cross section of the valley becomes narrower there. Accordingly, the stability of the subsiding layer is progressively reduced at lower levels, until it matches the neutral profile of the CBL underneath.

Figure 7 shows sequences of vertical θ profiles, where the distinction between different layers is clearly visible. As shown in the midvalley profiles, the sensible heat flux

FIG. 7. Vertical θ profiles taken every 30 min in the (top) narrow and (bottom) wide valleys, (left) at midvalley and (right) near the valley walls. The horizontal gray line marks the mountain-top height. Profiles in the right column are taken along a vertical line above the mountain-top level and along the slope below (i.e., they are pseudovetical below crest height). In all frames, profiles evolve in time from the unperturbed stable profile to the left (light gray) to the final well-mixed condition to the right (black).

at the valley floor first favors a localized increase in θ at the surface, while subsidence causes a downward displacement of the initial stable profile. The potential temperature gradient in the sinking stable layer is initially maintained, but it starts decreasing at lower levels as the compensation flow accelerates. In the wide valley, a well-mixed convective boundary layer with a nearly neutral thermal profile starts developing around 1400 LST. Heat advection and persisting turbulent mixing around crest height also cause the formation of an elevated mixed layer. This layer and the underlying CBL, initially separated by a stable region (a remnant of the stable valley core), progressively merge with each other.

The formation of an elevated weakly turbulent mixed layer and its progressive merging with a CBL growing underneath is similar to processes occurring in the vicinity of isolated slopes, observed by Reuten et al. (2007) and modeled by Serafin and Zardi (2010).

The along-slope vertical profiles in Fig. 7 clarify that localized warming occurs at the mountain tops: since the slope angle levels out near the summit, the advective cooling effect of the upslope flow vanishes there, and the atmosphere warms up. The buoyancy thus gained causes the detachment of thermal plumes, which in turn favor the onset of an entrainment heat flux from the free atmosphere above the LNB, as clearly visible from the negative θ perturbations observed in the two right panels of Fig. 7. In contrast, entrainment is negligible at midvalley, as shown in the two left panels.

Finally, Fig. 7 shows that the depth of the volume of atmosphere perturbed by the heating cycle is larger in the narrow valley (up to ~ 3800 m versus ~ 3300 m in the wide valley). Furthermore, the mixed layer temperature at the end of the heating phase is larger there (~ 309.1 K versus ~ 307.8 K in the wide valley), possibly as a consequence of

topographic effects (Whiteman 1990; Schmidli and Rotunno 2010).

4. Discussion

In this section we concentrate on understanding the factors that control the depth of the boundary layer in mountainous areas, and we analyze the relationship between warming due to advection related to slope flows and warming due to turbulent convection. Our analysis is supported by a study of the trajectories and thermal evolution of some selected air parcels.

a. Top-down and bottom-up warming

The flow pattern described in section 3b produces a “top-down” warming process. Heat is initially transferred to the atmosphere through sensible heat flux at the ground, but the heat imparted to the upslope flow ultimately reaches the valley core from above (i.e., through midvalley subsidence). Diabatic heating provided to the shallow mixed layer that grows over the valley floor can instead be regarded as a “bottom-up” warming process.

The analysis carried out so far—in particular, the observation that weaker upslope flow and mountain-top plumes occur when turbulent mixing is stronger—suggests that top-down and bottom-up heating are actually two competing processes, the former being most effective in narrow valleys and in the early phase of the diurnal cycle. This is shown in Fig. 8, where the time evolution of the leading terms in the potential temperature budget is represented. It is easy to see that the processes of subsidence warming (closely related to the occurrence of slope flows, and represented by $-w\partial\theta/\partial z$) and of turbulent heat diffusion (mostly related to the vertical gradient of the turbulent flux, $-\partial w'\theta'/\partial z$) are almost mutually exclusive: where one occurs, the other is negligible.

The interplay between the two mechanisms is represented in Fig. 9: upslope flow (labeled “1”) and thermal plumes at the crests (“2”) are compensated by horizontal flow toward the valley center feeding heat into a slowly destabilizing layer above the valley core (“3”), and by midvalley subsidence along the vertical (“4”). The region of downward motion in the valley core is slowly eroded by the CBL growing underneath (“5”) and becomes progressively less stable because of the downward vertical acceleration dynamically induced by the narrowing valley walls.

While the sketch in Fig. 9 offers an instantaneous snapshot of the flow field, Fig. 10, representing time–height series of the cross-sectional mean potential temperature, helps in understanding how the phenomenon evolves in time, affecting the thermal structure of the valley atmosphere. The downward bending of isentropes

in Fig. 10 indicates progressive warming. The valley atmosphere starts being heated approximately at 1100 LST, initially between 2000 and 3000 m, exactly where detrainment from the thermal plumes over mountain tops occurs. At that height, warming occurs in conjunction with a decrease in the static stability (diagnosed by increased spacing between isentropes).

At the same time, but at lower heights, midvalley subsidence brings potentially warm air toward the valley bottom. In a first phase, until approximately 1500 LST, the heating rate (evaluated from the inclination of isentropes $\partial\theta/\partial t$) is approximately equal at all heights below crest level, and it is larger in the narrower valley, where subsidence in the valley core is stronger.

After 1500 LST, temperature starts increasing at a lower rate in both valleys. Isentropes are now approximately vertical near the surface, especially in the wider valley. This indicates that a mixed (or only slightly stable) layer is being warmed uniformly along the vertical. The presence of a mixed layer is apparent from 1200 LST in the wide valley and from 1400 LST in the narrow one. These timings coincide with the moments at which upslope flow weakens, as explained in section 3b. The fact that turbulent mixing starts being efficient at 1400 LST in the narrow valley and at 1200 LST in the wide one is also confirmed by Fig. 8, where the intensity of the two phenomena is visually displayed.

As a concluding remark, note that the interplay between advective top-down warming and turbulent bottom-up warming in a real valley could be considerably altered during the afternoon phase by the presence of an upvalley wind, not included in our simulations. In particular, the initiation of upvalley flow would likely cause the onset of persistent cool horizontal advection and the enhancement of the shear production of turbulence, two processes with a potentially large impact on the dynamics of the atmospheric boundary layer.

b. Depth of the valley boundary layer

The simulations presented above suggest that slope flow circulations perturb the atmosphere of a valley even above the level of mountain tops, since thermal plumes developing there reach their LNB a few hundred meters above the crests.

Below LNB height, the atmosphere is subject both to advective heat transfer (cooling along the slopes and warming in the valley core) and to turbulent heat diffusion (near the valley floor). Both the CBL growth above the valley floor and the onset of upslope flows along the valley sidewalls can be regarded as boundary layer processes: in both cases the flow field is directly produced by the surface heat flux, is generally turbulent, and responds to rapid variations in surface forcing.

FIG. 8. Time–height series of the θ budget in the (left) narrow and (right) wide valleys. In the midvalley vertical profile, isolines are every $1.5 \times 10^{-4} \text{ K s}^{-1}$; positive and negative contours are black and gray, respectively. The horizontal gray line marks the mountain-top height.

Although nonturbulent in general, the region affected by compensating subsidence can also be considered a part of the boundary layer according to Stull's (1988) approach, which defines boundary layer processes as the near-surface phenomena responding to surface forcings on a time scale of around 1 h or less. Therefore, we refer to the volume of atmosphere perturbed by the onset of slope circulations as the *valley boundary layer* (VBL), by analogy to the concept of the mountain boundary layer recently used by De Wekker et al. (1998, 2002).

The upper limit of the VBL roughly corresponds to the top of the mixed or slightly stable layer produced by

the detrainment and horizontal advection of warm air from mountain-top plumes. The depth of the VBL can be estimated using the approach illustrated in Fig. 10. The thick black line represents the level of neutral buoyancy z_{lnb} of parcels in the mountain-top plumes and has been evaluated as follows:

$$z_{\text{lnb}} = z_{\text{top}} + \frac{\theta_{\text{top}} - \theta_{\text{top}}^0}{\Gamma}, \quad (2)$$

where $z_{\text{top}} = 2000 \text{ m}$ is the mountain-top height, θ_{top} is the surface potential temperature at mountain top, θ_{top}^0

FIG. 9. An idealized sketch of the circulation induced by upslope flows. Areas affected by upslope flow, thermal plumes, horizontal motion at mountain-top level, adiabatic subsidence, and turbulent convection are labeled 1–5, respectively. Arrows and whirls suggest the main features of the flow field. Asymmetries in the sketch reflect the structure of turbulent boundary layer eddies.

is the initial unperturbed value of the same parameter, and $\Gamma = 3 \times 10^{-3} \text{ K m}^{-1}$ is the thermal gradient of the undisturbed atmosphere. Equation (2) is merely a parcel-method estimation of the CBL depth, adapted to valley topography.

The thin black line in Fig. 10 defines the region subject to potential warming. Since it overlaps almost exactly with z_{lnb} , it is easy to see that the latter is a good estimate of the depth of the VBL during daytime.

The dotted line identifies instead a region subject to potential cooling; this part of the atmosphere [the entrainment layer (EL)] loses heat by entrainment toward the valley atmosphere. Indeed, since cooling is visibly

very limited in this region (note that isentropes remain almost horizontal), this contribution to the warming of the valley atmosphere is actually negligible. Moreover, the larger the valley, the smaller the contribution of the entrainment heat flux to its heat budget: in fact, entrainment is only significant in the immediate vicinity of mountain-top plumes, which become weaker and separated by increasing distances as the valley cross section becomes larger.

c. Lagrangian analysis

Tracking the properties of air parcels along their motion through the valley atmosphere further elucidates the mechanisms we described. Several tens of trajectories have been produced and analyzed using the trajectory computation capabilities of the ARPS model: here we only present three of them, related to results from the narrow valley run.

The first trajectory (Fig. 11) refers to an air parcel initially residing on the valley floor. The parcel experiences little heating for the first 2 h, when it does not move. It suddenly starts rising along a slope, and during the ascent it undergoes strong diabatic heating: its temperature rises by approximately 7 K in 30 min (i.e., the time it takes for the parcel to run along the 4000-m slope at a mean speed of 2.2 m s^{-1}). Once the mountain top is reached, the parcel ascends vertically, and its motion immediately becomes adiabatic. Since its temperature is 307.5 K at mountain-top level, the parcel should reach its LNB at 2500 m, according to Eq. (2). This is what actually happens, after an initial overshoot up to 2800 m. At the top of the plume and along the downdraft, the parcel is subject to mixing with the potentially warmer air in the EL and

FIG. 10. Time-height series of cross-valley mean θ in the (left) narrow and (right) wide valleys. Isolines are every 0.5 K, thick every 3 K; 315 K at top. The thin solid and dotted black lines mark regions of warming and cooling over time, respectively. The thick black line indicates an estimation of the valley boundary layer depth according to Eq. (2). The horizontal gray line marks the mountain-top height.

FIG. 11. Trajectory coordinates and thermal evolution of a parcel initially located at the valley bottom. The gray line in the lower left panel refers to the y coordinate. The initial ($\times$), final ($\star$), and intermediate positions every hour ($\bullet$) are marked. Hour 0 corresponds to 0900 LST.

thus further experiences little heating. From now on, the motion of the parcel is adiabatic, as the skew T - $\log p$ plot shows: at about hour 3, the parcel starts moving horizontally at LNB height toward the valley center; when this is reached, the motion becomes downward and approximately vertical. It takes about 6 h to reach the proximity of the valley floor again. The path covered by the parcel is essentially two-dimensional, since displacements along the longitudinal axis of the valley are very small.

The second trajectory (Fig. 12) clarifies that a parcel originally located in the valley core starts descending adiabatically immediately after the onset of the upslope flow. Hence, it contributes to warming the valley atmosphere through adiabatic compression. When the parcel reaches the valley bottom, it travels along a path similar to the first trajectory. Having a slightly higher temperature when it reaches the mountain top (308 K), it is pushed to higher altitudes compared to the previous one.

The third trajectory (Fig. 13) shows that actually things can be much more complicated: here a parcel,

initially at some height above the slope, is first captured by the compensation current of the anabatic wind and travels adiabatically downslope. Then it is captured in the growing upslope flow layer. While traveling upslope it gets heated, at least until it detrains and is captured by the downslope flow again: the motion again becomes adiabatic. Then the motion reverses once more, and the parcel finally reaches the mountain top and the thermal plume that stands there, in analogy to the trajectories analyzed above.

This complex sequence of events confirms that the upslope flow, which initially blows with constant intensity, turns into a succession of upslope moving thermals as soon as the available buoyancy exceeds the amount necessary to drive the upslope motion.

5. Conclusions

By means of idealized LES, we examined the flow field and heating processes that occur in two-dimensional mountain valleys during the diurnal phase of the daily cycle.

FIG. 12. As in Fig. 11, but for a parcel initially located 1500 m above the valley floor.

We highlighted that a peculiar indirect heating process occurs during the morning phase: thermal energy supplied at the surface is first transported upward along the slopes, up to the mountain tops and beyond. Warm air parcels then detrain laterally from thermal plumes above mountain tops and are captured by subsiding motions that compensate the upslope flow. This way they re-enter the valley atmosphere from above and contribute to its warming. Therefore, the free atmosphere descent mechanism first described by Whiteman and McKee (1982) appears to be limited to the atmospheric depth below the level of neutral buoyancy of the plumes developing on mountain tops.

We referred to the subsidence warming mechanism as a *top-down warming*, as opposed to *bottom-up warming* resulting from turbulent convection originating at the valley floor. We demonstrated that advective top-down warming and convective bottom-up warming antagonize each other, and that the former prevails in the morning whereas the latter is more efficient in the afternoon.

Note that the present idealized study, considering a topography allowing only for the generation of cross-valley air motions, was designed to be representative of

the dynamics of real valleys during the morning phase, when slope winds are the dominant flow pattern. As a consequence, our statements about the relevance of subsidence warming in the early diurnal phase are expected to maintain general validity.

Conversely, the suggested greater importance of turbulent heat transfer during the afternoon phase should be further investigated, considering in particular any possible effect related to the occurrence of along-valley breezes, often the dominant flow feature in the late afternoon. Relevant consequences of the presence of upvalley flow might include on one hand an increased shear production of turbulence (likely contributing to a further deepening of the convective boundary layer) and on the other hand a persistent advection of relatively cool air (possibly causing a contrary effect).

Through the simulations presented in this study, we have highlighted that the atmospheric boundary layer appears to have peculiar features in mountainous regions, essentially related to the presence of slope winds. The valley boundary layer (VBL) may include a CBL but is not limited to it. Elevated layers, where the atmosphere is warmed and becomes weakly turbulent due

FIG. 13. As in Fig. 11, but for a parcel initially located above a slope.

to advection of heat and TKE, are also included in it. A nonturbulent stable region where air subsides toward the valley floor to compensate upward motion along the slopes is also embedded in the VBL, at least in the early stages of its development. As the daily heating cycle progresses, the CBL and elevated stable layers may merge into a unique mixed layer. These features are consistent with previous findings from airborne measurements (de Franceschi et al. 2003; Rampanelli and Zardi 2004; Weigel and Rotach 2004).

The presence of moisture, neglected in the present study, will play a complex role in conditioning the efficiency of the valley heating process, causing the release of latent heat but also reducing the radiant energy flux if condensation occurs. Future investigations are needed to clarify its possible impact.

As a final remark, it has to be stressed once more that our simulations refer to a two-dimensional topographic setting: as such, they cannot reproduce any effect resulting from longitudinal thermal imbalances along a valley. For instance, the contrast between a VBL and a CBL developing above a plain would cause the onset of an upvalley wind, a phenomenon that cannot be appreciated

in the present model setup. However, our analysis of the properties of the VBL, once compared to the typical evolution of the fair-weather CBL over flat terrain, will provide insight on how the valley–plain thermal contrast, driving upvalley flow, is originally generated.

Acknowledgments. The authors acknowledge support from CINECA, the Italian national supercomputing centre, where the computations for this study were performed, and from the center CUDAM at the University of Trento.

REFERENCES

- Andretta, M., A. Weiss, N. Kljun, and M. W. Rotach, 2001: Near-surface turbulent momentum flux in an Alpine valley: Observational results. *MAP Newsletter*, No. 15, Meteo Swiss, Zurich, Switzerland, 122–125.
- Beljaars, A. C. M., 1994: The parametrization of surface fluxes in large-scale models under free convection. *Quart. J. Roy. Meteor. Soc.*, **121**, 255–270.
- Chow, F. K., A. P. Weigel, R. L. Street, M. W. Rotach, and M. Xue, 2006: High-resolution large-eddy simulations of flow in a steep alpine valley. Part I: Methodology, verification and sensitivity experiments. *J. Appl. Meteor. Climatol.*, **45**, 63–86.

- Defant, F., 1951: Local winds. *Compendium of Meteorology*, T. F. Malone, Ed., Amer. Meteor. Soc., 655–672.
- de Franceschi, M., G. Rampanelli, D. Sguerso, D. Zardi, and P. Zatelli, 2003: Development of a measurement platform on a light airplane and analysis of airborne measurements in the atmospheric boundary layer. *Ann. Geophys.*, **46**, 269–283.
- , D. Zardi, M. Tagliazucca, and F. Tampieri, 2009: Analysis of second-order moments in the surface layer turbulence in an Alpine valley. *Quart. J. Roy. Meteor. Soc.*, **135**, 1750–1765.
- De Wekker, S. F., 2002: Structure and morphology of the convective boundary layer in mountainous terrain. Ph.D. thesis, Department of Earth and Ocean Sciences, University of British Columbia, 191 pp.
- , 2008: Observational and numerical evidence of depressed convective boundary layer height near a mountain base. *J. Appl. Meteor. Climatol.*, **47**, 1017–1026.
- , S. Zhong, J. D. Fast, and C. D. Whiteman, 1998: A numerical study of the thermally driven plain-to-basin wind over idealized basin topographies. *J. Appl. Meteor.*, **37**, 606–622.
- , D. Steyn, M. Rotach, J. Fast, and S. Zhong, 2002: Observations and numerical modeling of the daytime boundary layer structure in the Riviera Valley, Switzerland. *Extended Abstracts, 10th Conf. on Mountain Meteorology*, Park City, UT, Amer. Meteor. Soc., 3.5. [Available online at http://ams.confex.com/ams/10Mountain/techprogram/paper_39536.htm.]
- Haiden, T., 2003: On the pressure field in the slope wind layer. *J. Atmos. Sci.*, **60**, 1632–1635.
- Kondo, J., T. Kuwagata, and S. Haginoya, 1989: Heat budget analysis of nocturnal cooling and daytime heating in a basin. *J. Atmos. Sci.*, **46**, 2917–2933.
- Moeng, C.-H., 1984: A large eddy simulation model for the study of planetary boundary layer turbulence. *J. Atmos. Sci.*, **41**, 2052–2061.
- Noilhan, J., and S. Planton, 1989: A simple parameterization of land surface processes for meteorological models. *Mon. Wea. Rev.*, **117**, 536–549.
- Oke, T. R., 1987: *Boundary Layer Climates*. Routledge, 435 pp.
- Prandtl, L., 1952: Mountain and valley winds in stratified air. *Essentials of Fluid Dynamics*, Hafner, 422–425.
- Rampanelli, G., and D. Zardi, 2004: A method to determine the capping inversion of the convective boundary layer. *J. Appl. Meteor.*, **43**, 925–933.
- , —, and R. Rotunno, 2004: Mechanisms of up-valley winds. *J. Atmos. Sci.*, **61**, 3097–3111.
- Ren, D., and M. Xue, 2004: A revised force-restore model for land surface modeling. *J. Appl. Meteor.*, **43**, 1768–1782.
- Reuten, C., D. G. Steyn, and S. E. Allen, 2007: Water tank studies of atmospheric boundary layer structure and air pollution transport in upslope flow systems. *J. Geophys. Res.*, **112**, D11114, doi:10.1029/2006JD008045.
- Rotach, M. W., and D. Zardi, 2007: On the boundary layer structure over complex terrain: Key findings from MAP. *Quart. J. Roy. Meteor. Soc.*, **133**, 937–948.
- Schmidli, J., and R. Rotunno, 2010: Mechanisms of along-valley winds and heat exchange over mountainous terrain. *J. Atmos. Sci.*, **67**, 3033–3047.
- Schumann, U., 1990: Large-eddy simulation of the up-slope boundary layer. *Quart. J. Roy. Meteor. Soc.*, **116**, 637–670.
- Serafin, S., and D. Zardi, 2010: Structure of the atmospheric boundary layer in the vicinity of a developing upslope flow system: A numerical model study. *J. Atmos. Sci.*, **67**, 1171–1185.
- Steinacker, R., 1984: Area–height distribution of a valley and its relation to the valley wind. *Beitr. Phys. Atmos.*, **57**, 64–71.
- Stull, R. B., 1988: *An Introduction to Boundary-Layer Meteorology*. Kluwer, 670 pp.
- Vergeiner, I., and E. Dreiseitl, 1987: Valley winds and slope winds—Observations and elementary thoughts. *Meteor. Atmos. Phys.*, **36**, 264–286.
- Weigel, A. P., and M. W. Rotach, 2004: Flow structure and turbulence characteristics of the daytime atmosphere in a steep and narrow Alpine valley. *Quart. J. Roy. Meteor. Soc.*, **130**, 2605–2627.
- , F. K. Chow, M. W. Rotach, R. L. Street, and M. Xue, 2006: High-resolution large-eddy simulations of flow in a steep alpine valley. Part II: Flow structure and heat budgets. *J. Appl. Meteor. Climatol.*, **45**, 87–107.
- Wenger, R., 1923: Zur Theorie der Berg- und Talwindes. *Meteor. Z.*, **40**, 193–204.
- Whiteman, C. D., 1982: Breakup of temperature inversions in deep mountain valleys. Part I: Observations. *J. Appl. Meteor.*, **21**, 270–289.
- , 1990: Observations of thermally developed wind systems in mountainous terrain. *Atmospheric Processes over Complex Terrain*, Meteor. Monogr., No. 45, Amer. Meteor. Soc., 5–42.
- , 2000: *Mountain Meteorology: Fundamentals and Applications*. Oxford University Press, 355 pp.
- , and T. B. McKee, 1982: Breakup of temperature inversions in deep mountain valleys. Part II: Thermodynamic model. *J. Appl. Meteor.*, **21**, 290–302.
- Xue, M., K. K. Droegemeier, and V. Wong, 2000: The Advanced Regional Prediction System (ARPS): A multi-scale non-hydrostatic atmospheric simulation and prediction model. Part I: Model dynamics and verification. *Meteor. Atmos. Phys.*, **75**, 161–193.
- , and Coauthors, 2001: The Advanced Regional Prediction System (ARPS): A multi-scale nonhydrostatic atmospheric simulation and prediction tool. Part II: Model physics and applications. *Meteor. Atmos. Phys.*, **76**, 143–165.