The Impact of the Temperature Inversion Breakup on the Exchange of Heat and Mass in an Idealized Valley: Sensitivity to the Radiative Forcing

DANIEL LEUKAUF, ALEXANDER GOHM, MATHIAS W. ROTACH, AND JOHANNES S. WAGNER*
Institute of Atmospheric and Cryospheric Sciences, University of Innsbruck, Innsbruck, Austria

(Manuscript received 27 March 2015, in final form 13 July 2015)

ABSTRACT

The breakup of a nocturnal temperature inversion during daytime is studied in an idealized valley by means of high-resolution numerical simulations. Vertical fluxes of heat and mass are strongly reduced as long as an inversion is present; hence it is important to understand the mechanisms leading to its removal. In this study breakup times are determined as a function of the radiative forcing. Further, the effect of the nocturnal inversion on the vertical exchange of heat and mass is quantified. The Weather Research and Forecasting Model is applied to an idealized quasi-two-dimensional valley. The net shortwave radiation is specified by a sine function with amplitudes between 150 and 850 W m\(^{-2}\) during daytime and at zero during the night. The valley inversion is eroded within 5 h for the strongest forcing. A minimal amplitude of 450 W m\(^{-2}\) is required to reach the breakup, in which case the inversion is removed after 11 h. Depending on the forcing amplitude, between 10% and 57% of the energy provided by the surface sensible heat flux is exported out of the valley during the whole day. The ratio of exported energy to provided energy is approximately 1.6 times as large after the inversion is removed as before. More than 5 times the valley air mass is turned over in 12 h for the strongest forcing, whereas the mass is turned over only 1.3 times for 400 W m\(^{-2}\).

1. Introduction

The formation of cold-air pools in valleys and sinkholes during clear nights is a well-known process that has been intensively studied by means of field experiments (e.g., Clements et al. 2003; Steinacker et al. 2007; Price et al. 2011; Lareau et al. 2013) and numerical simulations (e.g., Zängl 2005; Burns and Chemel 2014; Vosper et al. 2014). The strength and depth of the temperature inversion that develops during nighttime strongly influence the development of the boundary layer during daytime (Schnitzhofer et al. 2009) and therefore also the exchange of heat, moisture, momentum, and pollutants with the free atmosphere above the valley. A key variable for the inversion-breakup (hereinafter shortened to breakup) process as well as for the exchange is the solar forcing, which has a direct impact on the sensible heat flux at the surface. The key goals of this study are to determine the dependence of the vertical heat and mass fluxes on the amplitude of the net solar radiation and to identify breakup times for these forcing conditions.

Thermally driven winds that form along slopes of mountains under clear-sky conditions provide an effective exchange mechanism. Henne et al. (2004) have studied these mechanisms over two deep Alpine valleys and estimated that under such conditions the total mass inside these valleys is recirculated up to three times in 24 h. It is also clear from previous studies that a completely mixed valley atmosphere is not a precondition to transport heat (Schmidli 2013), moisture (Weigel et al. 2007), and pollution (Lehner and Gohm 2010) out of the valley atmosphere. Upslope winds that form along heated terrain provide an effective transport mechanism, being able to penetrate stable layers. In the...
presence of strong elevated inversions, however, a splitting of the upslope flow may occur and consequently only a fraction of the flow penetrates the inversion while the rest is recirculated below, forming a secondary circulation (Vergeiner and Dreisittl 1987). As soon as the breakup is reached, the exchange may strongly increase as long as the breakup does not occur too late in the afternoon when the radiative forcing is already decreasing. Hence the time of the inversion breakup, which depends on the atmospheric stability as a function of height, the valley geometry, the soil moisture, and the forcing, is expected to be an important parameter that influences the bulk exchange of properties between the valley and the free atmosphere.

Whiteman (1982) has described three mechanisms that are responsible for the destruction of the stable boundary layer inside a valley: heating of the air near the valley bottom where the height of the inversion top remains constant, subsidence in the center of the valley causing the top of the inversion to sink, and a combination of the first two mechanisms. As described in Whiteman and McKee (1982), the breakup will be achieved if the energy provided to the valley atmosphere through the surface sensible heat flux is equal to or exceeds the energy required to remove the cold pool and achieve a neutral stratification inside the valley, assuming that no heat can leave the valley before breakup. If energy “leaks” out of the valley, however, it is likely that the breakup will be delayed and may never be reached.

The influence of the valley geometry and shading effects on the inversion breakup has been studied by Colette et al. (2003). They found inversion lifetimes of 4–7 h for forcing conditions equivalent to 21 September at the equator. The inversion persists longer in deeper valleys, and inversion lifetimes converge to the one over a plain as the width of the valley floor is increased. They came to the conclusion that shading in steep and narrow valleys delays the onset of the upslope winds but increases the breakup time by only up to 30 min. Idealized simulations of the destruction of the stable layer in a valley have been performed by Anquetin et al. (1998) for a winter case and a summer case. In both cases, however, the forcing was too weak to completely mix the valley atmosphere and a stable layer persisted in the upper part of the valley.

Despite the importance of thermally driven winds for the vertical exchange, they are not represented or are only partially represented in global numerical weather prediction and climate models, since these models have still resolutions that are too coarse to capture the complexity of mountainous terrain including individual valleys. This has an impact on the representation of the boundary layer over complex terrain in these models. During the night, strong surface inversions are not captured correctly, a problem that is reported even for mesoscale models with a resolution of a few kilometers (Vosper et al. 2013). In addition, the boundary layer schemes that are applied in today’s operational models are, in strict terms, only valid for flat and horizontally homogeneous terrain, and it cannot a priori be expected that they represent the boundary layer structure over mountains in a satisfying way (Rotach and Zardi 2007). The error caused by unresolved topography and turbulence processes has been investigated by Wagner et al. (2014). Large errors appear in the cross-valley circulation because of unresolved topography, especially if the valley depth is not represented correctly and boundary layers are too warm during daytime if the terrain is overly smoothed. The fact that exchange processes are neither explicitly resolved nor parameterized leads to errors that also influence the mean state of the free atmosphere above the valley.

To gain insight to the boundary layer processes that affect exchange with the free atmosphere, field experiments (e.g., Weigel and Rotach 2004; Rotach and Zardi 2007), numerical simulations (Chow et al. 2006; Weigel et al. 2006), and analytical studies (e.g., Noppel and Fiedler 2002) have been conducted. The importance of the valley geometry and the plain-to-mountain circulation for vertical transport processes has recently been highlighted by Wagner et al. (2015b). They showed that air parcels over the foreland are advected by the up-valley wind and rise with the slope winds to altitudes far above crest height. The return flow over the valley brings these parcels back to the foreland at higher altitudes than the boundary layer height over the plain. The described mechanism takes place in every valley of a larger mountain range and forms the phenomenon known as alpine pumping, in which a “heat low” over the mountain range causes the air to rise and be replaced by an inflow from the foreland. By this mechanism, the vertical exchange is increased by a factor of 3–4 in comparison with that of a plain. Narrow and deep valleys produce a stronger circulation than wide and shallow valleys. Valleys with a rising valley floor or narrowing valley produce stronger valley winds as well, increasing the strength of the plain-to-mountain circulation (Wagner et al. 2015a). These studies have emphasized the need for a representation of vertical-exchange processes in models with a coarse resolution (Rotach et al. 2014).

There is a wealth of literature on idealized numerical simulations that address thermally driven flows (e.g., Serafin and Zardi 2010a,b; Schmidli and Rotunno 2010; Wagner et al. 2014, 2015b), but in these studies the
surface sensible heat flux or radiative forcing is either constant or is chosen to represent a certain day of the year and latitude. This limitation calls for a systematic study to quantify the dependence of the vertical exchange on the net radiative forcing. In this paper, we vary the forcing amplitude systematically to find a relation between the forcing and the inversion lifetime. This approach also allows us to identify the minimal forcing amplitude required to remove a valley inversion and to determine the amount of heat exported during and after the temperature inversion breakup. Furthermore, we want to quantify the mass flux out of the valley to estimate the effectiveness of the venting mechanism as a function of the forcing amplitude. To reduce the number of free parameters, we restrict this study to a simple straight valley and to inversions that develop during the first night of model integration.

2. Methods

a. Model setup and configuration

In this study we use the Advanced Research Weather Research and Forecasting (WRF) Model, version 3.4, which has already been successfully applied for idealized simulations of thermally driven flows, both at kilometer resolution and for large-eddy simulations (e.g., Catalano and Moeng 2010; Schmidli et al. 2011; Wagner et al. 2014, 2015a,b). The nonhydrostatic, fully compressible WRF Model uses a horizontally staggered Arakawa-C grid and a terrain-following pressure vertical coordinate. In this study, we run it without a planetary boundary layer scheme but use a 1.5-order three-dimensional turbulence kinetic energy closure (Deardorff 1980; Skamarock et al. 2008) for the subgrid-scale turbulence parameterization. The integration is done with a third-order Runge–Kutta scheme in time and with a fifth-order (third-order) scheme for the horizontal (vertical) advection. For the surface layer, the “MM5” similarity scheme with Monin–Obukhov similarity functions (Dyer and Hicks 1970; Paulson 1970; Webb 1970) and four stability regimes (Zhang and Anthes 1982) is used and land surface processes are represented by the “Noah” land surface model (Niu et al. 2011) with four soil layers. For moist processes, the WRF single-moment 6-class scheme (Hong and Lim 2006) is used, although the initial conditions are designed to avoid the formation of clouds most of the time. Still, shallow cumulus clouds form sporadically, but the influence of the latent heat release on the air temperature is negligible. Coriolis effects are neglected in this study, which is appropriate, since slope flows have a high Rossby number.

The topography used in this study consists of two sine-shaped slopes with a peak height of 1.5 km and a 1-km-wide horizontal valley floor (Fig. 1). This is similar to the geometry of the “REF2D” simulation of Schmidli (2013) except that the domain borders are located at the mountain crests in that study, leading to a 20-km-wide domain. The influence of the valley surroundings on slope and valley winds has been studied by Schmidli and Rotunno (2012), who found only small differences between two mountains surrounded by a plain and an infinite train of valleys and ridges. Hence the chosen terrain boundaries are not expected to be critical. A valley length of 20 km is used to enable the formation of a sufficient number of convective cells to allow a meaningful calculation of mean and resolved turbulent fluxes. Since the geometry is quasi 2D, no valley winds but only slope winds can form.

With a horizontal mesh size of 200 m, the domain has 101 × 101 data points in the horizontal plane. In the vertical direction, we are using 109 levels with vertical stretching of the level distance. In the valley center, the lowest model level lies 8.1 m above the surface and the distance between the lowest two model levels is 8.51 m. The model top lies 12.2 km above sea level. A horizontal resolution on the order of 200 m has been applied successfully by various authors (i.e., Colette et al. 2003; Weigel et al. 2007; Wagner et al. 2015b) for simulations of daytime slope winds. Test simulations (not shown) with horizontal mesh sizes of 50 and 100 m, respectively, show that mean variables such as the average potential temperature and bulk vertical heat fluxes are very similar. Hence the chosen resolution is justified for the aims of this study.

The lateral boundary conditions are periodic in both the x and y directions, leading to an infinitely long train of parallel valleys. The upper 4.7 km of the domain are
occupied by a Rayleigh damping layer with a damping coefficient of 0.003 s\(^{-1}\).

The model is initialized using a sounding with a constant Brunt–Väisälä frequency of \(N = 0.01 \text{s}^{-1}\), a surface temperature \(T_0 = 280\, \text{K}\), a surface pressure \(p_0 = 1000\, \text{hPa}\) at \(z = 0\, \text{m}\) MSL, and 40% relative humidity throughout the whole atmosphere. The atmosphere is initially at rest. The soil type “sandy loam” is used, and soil moisture is initialized with 20% of the saturation value. This condition leads to a latent heat flux that is virtually zero and, in combination with the 40% relative humidity, causes practically cloud free conditions during the whole simulation period. The vegetation type is semidesert with a leaf area index of 1.5, a vegetation fraction of 0.1, and a surface roughness length of 0.16 m. These initial and surface conditions are homogeneous everywhere, similar to those in Schmidli and Rotunno (2010). A randomly distributed potential temperature perturbation of 0.5 K is initially added to the base state at the surface and at the lowest five model levels to trigger convection.

b. Forcing

The net shortwave radiative flux through a horizontal plane at the surface is defined as

\[
F_{\text{sw}} = \begin{cases} 
A_{\text{sw}} \sin \left(\frac{2\pi(t - 6)}{24}\right) \cos(\alpha) & 6 < t < 18 \\
0 & \text{otherwise}
\end{cases}.
\]  

(1)

Here, \(A_{\text{sw}}\) is the net shortwave amplitude and \(\alpha\) is the slope angle, which is about 15° at the steepest point of the chosen terrain. The \(t\) limits of 6 and 18 correspond to 0600 and 1800 LT, respectively. This definition of the shortwave radiation would, strictly speaking, be appropriate for a valley that is located at the equator and oriented in an east–west direction. An asymmetric forcing of the valley slopes is avoided this way. Because of the moderate slope angle, the forcing amplitude at the slopes is reduced by only \(\sim 3.3\%\) at the steepest point of the slope relative to a flat surface.

The definition in Eq. (1) has the advantage that basically only the amplitude \(A_{\text{sw}}\) defines the net shortwave radiation since the influence of \(\alpha\) is negligible, whereas in the case of a more sophisticated scheme the date, latitude, and albedo would have to be specified. With this simplified approach, we neglect the effects of absorption, scattering, and emission of radiation on the potential temperature tendencies. Only the surface is heated by the radiation, which then leads to the atmosphere being heated by turbulent heat fluxes. Simulations for 15 different amplitudes are carried out, varying \(A_{\text{sw}}\) from 150 to 850 W m\(^{-2}\) by increments of 50 W m\(^{-2}\). The chosen range of forcing amplitudes is representative for a net solar radiation that can be observed at an equinox within approximately 80°S and 80°N with an albedo of 27%. Hence with this choice we cover a broad range of radiation conditions that are typical for many parts of the world. These 15 simulations are labeled A150–A850 in the following, according to the respective amplitude of net shortwave radiation.

The upwelling longwave flux at the surface is parameterized using \(F_{\text{lw}}^1 = \varepsilon a T_{\text{sk}}^4\), where \(\varepsilon = 0.92\) and \(T_{\text{sk}}\) is the skin temperature. For the downwelling longwave flux at the surface, we use the empirical formula by Angström (2013):

\[
F_{\text{lw}}^{1} = \sigma T^4(a - b \times 10^{-cc}),
\]  

(2)

where \(\varepsilon\) is the partial pressure of the water vapor and the coefficients \(a = 0.790, b = 0.174,\) and \(c = 0.055\, \text{hPa}^{-1}\). Although the temperature \(T\) is originally taken 2 m above ground, we decided to use the temperature at the fourth model level (\(\sim 35\, \text{m}\)) instead, since using the 2-m temperature leads to too-strong diurnal variations. With this definition, the downwelling longwave flux of the A850 simulation shows differences of less than 4% relative to the flux of a simulation using the Rapid Radiative Transfer Model for GCMs (Clough et al. 2005), placed at the equator and started at 21 March.

The model is initialized at 0600 LT and is run for 60 h. During the first 24 h, the model develops a convective boundary layer during the day and a cold-air pool during the night. With this approach, we create appropriate initial conditions for studying the breakup of the temperature inversion on the second day of the simulation. Unless stated otherwise, data in the following sections are always from the second day. The evolution of the atmosphere during the first day is regarded as spinup. Data from the third day are used to assess whether the solution is periodic, which would imply that processes on the third day are essentially the same as on the second day.

It must be emphasized that it is the surface sensible heat flux that heats the valley that requires a spinup to become periodic, that is, to repeat on each consecutive day. The surface sensible heat flux of the second day is smaller than on the first day because of a larger ground heat flux. This is true for all simulations but is especially relevant for simulations with a weak forcing. The amplitude of the sensible heat flux increases linearly with solar forcing amplitude \(A_{\text{sw}}\), however, which fits the needs of this study. A linear regression of the amplitude of the horizontally averaged surface sensible heat flux \(A_{\text{shf}}\) with \(A_{\text{sw}}\) in the form \(A_{\text{shf}} = kA_{\text{sw}} + d\) gives \(k = 0.649\) and \(d = -91.7\, \text{W m}^{-2}\) for the first day and \(k = 0.643\) and
\( d = -104.2 \text{ W m}^{-2} \) for the second day. Hence our selected range of \( A_{sw} \) corresponds to a range of surface sensible heat flux amplitudes of \( \sim 5-450 \text{ W m}^{-2} \), which covers observed values in real-case studies (e.g., Rotach and Zardi 2007).

c. Averaging methods

The averaging methods used in this study follow the approach of Schmidli (2013) and are similar to those in Wagner et al. (2014). For convenience, it is repeated here that a turbulent flow variable \( \tilde{\psi} = \tilde{\psi} + \psi' \) is separable into the gridbox average \( \bar{\psi} \) and the subgrid-scale perturbation \( \psi' \), which has to be parameterized. The resolved turbulent part of a variable is \( \psi'' = \bar{\psi} - \langle \bar{\psi} \rangle \), where the averaging operator \( \langle \rangle \) is defined as

\[
\langle \tilde{\psi}(x, t) \rangle = \frac{1}{TL_y} \int_{t-\frac{T}{2}}^{t+\frac{T}{2}} \int_0^{T} \tilde{\psi}(x, y^*, z^*, t^*) \, dy^* \, dt^*.
\]

Here, the averaging intervals are set to \( L_y = 20 \text{ km} \) and \( T = 40 \text{ min} \); the latter is the same as in Schmidli (2013) and Wagner et al. (2014). This leads to a complete decomposition of \( \tilde{\psi} \) into mean, resolved turbulent, and subgrid-scale parts:

\[
\tilde{\psi} = \langle \tilde{\psi} \rangle + \psi'' + \psi'.
\]

If a full horizontal average is required, the corresponding parameter \( f \) is interpolated vertically to a constant height level and averaged horizontally, which is then denoted as \( [f]_{xy} \).

Applying this procedure to \( \bar{\Theta} \) and \( \bar{w} \) gives the total (TOT) kinematic vertical heat flux and its components, the mean (MEA), resolved turbulent (RES), and subgrid-scale turbulent (SGS) terms, respectively:

\[
\langle \bar{w}(\bar{\Theta}) \rangle = \langle \bar{w}(\bar{\Theta}) \rangle + \langle \bar{w}(\psi') \rangle + \langle \bar{w}(\Theta') \rangle.
\]

The mean part of the vertical heat flux has hereby to be redefined as \( \langle \bar{w} \rangle \langle \bar{\Theta} \rangle \), with \( \langle \bar{\Theta} \rangle = \langle \bar{\Theta} \rangle - \Theta_0(z) \). Here, \( \Theta_0(z) \) is the horizontal average of \( \langle \bar{\Theta} \rangle \) at a given height. The reasoning for this approach can be found in Wagner et al. (2014) and Schmidli (2013).

3. The breakup of the stable layer

a. Flow regimes

During the first 12 h of simulation time, slope winds and a convective boundary layer develop. Close to sunset, convection ceases, the slopes begin to cool, and katabatic winds form along the slopes (not shown). During the night, a cold pool forms at the bottom of the valley, and hence a three-layer atmosphere is established that represents the initial stratification for the second day (Fig. 2). This profile consists of a cold-air pool in the lowest few hundred meters, a weakly stratified intermediate layer, and a layer that represents the free atmosphere aloft, which still has the stability with which the model was initialized. The initial stratification has an impact on the development of the valley boundary layer and will be discussed in detail below. The following analysis of the breakup process is based on cross sections of the cross-valley wind and potential temperature; both averaged along the valley and over 41 min centered around the analyzed time step. These transects are shown for the A400 and the A700 simulation in Fig. 3.

At 0600 LT, when the sun rises, downslope winds are still active and persist in most simulations until 0700 LT. As the slopes are heated during the following hour, the katabatic flow stops and calm conditions appear over the slopes. Figure 3a is an example that is valid for A400, and very similar situations appear for all simulations. In this regime, no vertical exchange of mass or heat occurs; that is, the valley is sealed (SEA). Simulations with a relatively weak forcing (\( \leq 500 \text{ W m}^{-2} \)) stay in this regime considerably longer than 2 h. In the extreme case of the A150 simulation, the downslope winds or practically calm conditions persist during the whole day.
As soon as the surface heat flux becomes sufficiently positive, the air above the ground is heated and the slope winds reverse to an upslope direction. The thickness of the slope-wind layer is much smaller inside the cold-air pool than in the weakly stratified upper part of the valley (Figs. 3b,c). This change in the layer thickness is associated with an incorporation of air from the weakly stratified layer that results in a cross-valley circulation. Although the slope winds are not yet fully developed, the valley begins to “leak” air, and exchange with the

Fig. 3. The evolution of the valley boundary layer during the second day at (a), (b) 0800, (c), (d) 1000, and (e), (f) 1200 LT for $A_{\infty} =$ (left) 400 and (right) 700 W m$^{-2}$. Shown are the horizontal cross-valley wind component (m s$^{-1}$; color contours), potential temperature (K; black contours with 1-K increment), and wind vectors. All variables have been averaged over 41 min and over the full domain in the along-valley direction. To improve the visibility, not all of the vectors are drawn. The regimes are SEA in (a), TRA in (b) and (c), SC in (d) and (e), and MIX in (f). These regimes are defined in Table 1.
free atmosphere takes place. We call this the transition (TRA) regime. The simulations from A200 to A850 stay between 2 and 3 h in this regime.

With increasing surface heat flux, an unstable layer forms at the surface, leading to the onset of convection and strong vertical mixing. A mixed layer develops at the bottom of the valley, which is topped by an elevated inversion. The top of the mixed layer is located at \( \sim 650 \) m for the A400 simulation at 1200 LT and at \( \sim 800 \) m for A700 at 1000 LT. The inversion is penetrated by the slope winds (Figs. 3d,e). Only the core of the valley retains a stable stratification; therefore this regime is called stable core (SC). Although the slope winds are stronger than before, the stable core still represents a limiting factor for vertical-exchange processes since the mass flux in the slope-wind layer is inversely proportional to the stability. The elevated inversion locally reduces the mass flux in the slope-wind layer, and one part of the flow from below cannot pass the inversion and is redirected toward the valley center to form a secondary circulation—a process that has already been described by Vergeiner and Dreiseitl (1987) and is visible in Figs. 3d and 3e.

As the last remnants of the stable core are removed and the valley atmosphere becomes mixed (MIX), a single cross-valley circulation emerges and no secondary circulation exists anymore (Fig. 3f). Since the vertical exchange is no longer hindered by any stable layers, considerably stronger vertical fluxes are expected than before the breakup. Simulations with a forcing amplitude of less than 450 W m\(^{-2}\) cannot reach this regime since the heating rate is too small to break up the cold pool completely.

In the simulations with forcing amplitudes of 600 W m\(^{-2}\) and more, an additional pattern appears in the valley cross sections in the afternoon (not shown). The main updraft is no longer located at the mountain crest, where the two slope flows from each side of the crest converge, but the two primary updrafts are shifted toward the valley center. Hence two additional recirculation cells are established in counterrotation to the prior ones, which leads to a downdraft over the mountain crests. This regime is denoted mixed–shifted (MIXs) and is shown in Fig. 4. The appearance of this circulation pattern is accompanied by a growth of the horizontal width of convective cells from \( \sim 1 \) km at 1100 LT to \( \sim 5 \) km at 1700 LT. This pattern is not caused by the periodic boundary conditions, since a test simulation with two mountain ridges flanked by a 40-km-broad foreland shows a similar behavior. The MIXs regime appears only in the case of a highly symmetric setup in the absence of a synoptic flow or asymmetric forcing. Two unequally heated slopes will alter the position of the convective plumes, and the pattern would be superimposed by another circulation. This regime is a variant of the MIX regime since it has similar characteristics. In both regimes, the valley is completely mixed and thermals rising from the valley center can leave the valley atmosphere.

As the radiative forcing decreases in the afternoon, the sign of the surface heat flux becomes eventually negative, the slope winds reverse to downslope directions, and the static stability of the valley increases. In such a late-afternoon regime, the stability of the valley atmosphere is still relatively small, but the associated vertical fluxes are very weak. This regime is called moderately stable (MS), in contrast to the SEA regime, which is also characterized by calm conditions or downslope winds but is also characterized by a stronger vertical stability. During the night, the MS regime transforms into the SEA regime and the daily cycle is complete.

The evolution of the valley atmosphere, divided into these six stages, is summarized in a regime diagram (Fig. 5) and in Table 1, where the formal conditions for the regimes are defined. The focus of the following analysis is on the discussion of the second day of simulation (Fig. 5b). For the sake of completeness we have added the regime diagram for the first day (Fig. 5a). It shows that the atmosphere starts in the MS regime and proceeds directly to the SC regime. This is due to the lower initial stability, which makes the occurrence of the SEA and TRA regimes impossible. The MIX regime occurs 1 h later and only for simulations with a forcing of 500 W m\(^{-2}\) or more. The reason for this difference is discussed in the following paragraphs in detail. The MIXs appears only in the A850 simulation and appears
The pattern shown in Fig. 3 strongly depend on the stratification at 0600 LT, which is different for each simulation (see Fig. 2). In general, a three-layer atmosphere forms, with a cold-air pool topped by an inversion, a weakly stratified intermediate layer, and an unaltered background layer.

There are two mechanisms leading to the formation of the intermediate layer. Simulations with forcing amplitudes of 500 W m$^{-2}$ or higher receive enough energy to completely mix the initially stably stratified valley atmosphere during day 1. In this case, the intermediate layer is essentially a residual layer. An elevated inversion at the top of the cold pool forms during the night, not only as a result of surface cooling but also of the cold downslope winds. As the stability and depth of the cold pool increase during the night, the downslope winds cannot reach the bottom of the valley anymore, since they become positively buoyant with respect to the conditions there and contribute to the cooling closer to the top. Hence a strong temperature gradient forms at the top of the cold pool. For the A700 simulation at 0400 LT, Fig. 6a shows a profile of the potential temperature and Fig. 6b shows the strong cooling by advection in the budget of the potential temperature tendencies at $\sim$500 m MSL. The second mechanism for the formation of an intermediate layer is dominant in the case of a forcing amplitude of lower than or equal to $\sim$400 W m$^{-2}$. The convection, which develops at the valley bottom during the day, is relatively weak, and so are the slope winds. The valley atmosphere is heated but not sufficiently to neutralize the initial stratification completely. Also, the slope winds that converge above the crest do not reach far up into the atmosphere. Hence the stability is reduced in only a few hundred meters above the crest (see Fig. 2).

The impact of the initial stratification on the evolution of the valley atmosphere during the day becomes clear by comparing the sequence of regimes during the first and second days for a given amplitude $A_{sw}$ (Fig. 5). On the first day, when all simulations start from the same initial profile at 0600 LT, similar regimes occur as on the second day. However, since the very stable stratification
below 500 m is missing, the atmosphere is initially in the MS regime and proceeds directly to the SC regime as slope winds develop. Because of the missing intermediate layer, more energy is required to reach the MIX regime; hence the breakup occurs 1 h later and only for simulations forced with $A_{sw} \geq 500 \text{ W m}^{-2}$. On the first day, the MIXs regime appears only in the A850 simulation at 1700 and 1800 LT.

c. The breakup time

The breakup time $t_b$ marks the transition from the SC to the MIX regime and is reached when the whole valley atmosphere becomes neutral. Elevated inversions cannot hinder transport processes anymore, and thermals rising from the valley bottom reach heights above the mountain crest. In addition, the thickness of the slope-wind layer is strongly enhanced (cf. Figs. 3d and 3f).

In this study, we define that the breakup is achieved if, in the center of the valley, at each model level below crest height, the squared Brunt–Väisälä frequency $N^2(z)$ is less than the threshold value of $2.5 \times 10^{-5} \text{ s}^{-2}$, which corresponds to a temperature gradient of approximately $0.93 \text{ K (100 m)}^{-1}$. For the present simulations, this breakup time coincides with the time when the average flow pattern changes from at least two vertically stacked circulations cells to a single cross-valley circulation, that is, from Fig. 3d to Fig. 3f.

The inversion lifetime is defined as $\tau = t_b - t_i$, where $t_i$ is the initial time, that is, 0600 LT of the corresponding day. Hence the inversion lifetime is the time that a given inversion persists while being heated. It is a function of the forcing amplitude and is shown in Fig. 7 for all three days of simulation. The breakup is reached ~1 h later on the first day than on subsequent days because of the, on average, higher stability of the initial stratification as discussed above, and the breakup times for the second and the third days are nearly identical. In fact, the whole evolution of the flow on the third day is very similar to that of the second day. Hence a quasi-periodic steady state has been achieved already after 2 days of

![Fig. 6. Vertical profiles of the along-valley averaged (a) potential temperature (K) and (b) heating-rate components (10^{-3} \text{ K s}^{-1}) taken at the valley center for the A700 simulation at 0400 LT. The components shown in (b) are the potential temperature tendencies due to advection (PTADV) and turbulence (PTTRB) as well as the sum of both (PTTND).]
simulations. A minimum forcing amplitude of 450 W m\(^{-2}\) is required on the second and the third day to reach the breakup, whereas a value of 500 W m\(^{-2}\) is required on the first day. In the case of the strongest forcing, the breakup is reached at 1200 LT on the first day and at 1100 LT on the second and third days.

Colette et al. (2003) defined a breakup time on the basis of the Pasquill–Gifford stability classes (Seinfeld and Pandis 1997), where potential temperature gradients from \(-5\) to 5 K km\(^{-1}\) are defined as “neutral.” A potential temperature gradient of 5 K km\(^{-1}\) corresponds to \(\Lambda^2 \approx 17.2 \times 10^{-5} \text{s}^{-2}\), which is considerably larger than the threshold we have found to be suitable. Using this threshold would lead to inversion lifetimes of approximately 1 h less. Further, a breakup would be diagnosed for forcing amplitudes as low as 300 W m\(^{-2}\). A closer inspection of such cases shows that the flow pattern does not agree with the typical single circulation cell of the mixed regime (Fig. 3f) and that a weak elevated inversion is still present in the valley.

4. The vertical exchange of heat

In this section we will examine how much energy is required to remove the cold pool, how long it takes until it is removed, and how much energy is exported out of the valley. If the synoptic conditions are weak, the breakup depends solely on the surface sensible heat flux \(H_s(t)\), the profile of the potential temperature \(\Theta(z)\) at sunrise, and the valley geometry. As described in Whiteman and McKee (1982), the breakup will be achieved if the provided energy \(Q_{\text{prov}}\) is equal to the required energy \(Q_{\text{req}}\), that is,

\[
Q_{\text{prov}} = Q_{\text{req}} \quad \text{at} \quad t = t_b.
\]

In more specific terms,

\[
Q_{\text{prov}} = A \int_{t_i}^{t_b} H_s(t) \; dt
\]

is the total energy provided through surface area \(A\) from sunrise until breakup time \(t_b\) and

\[
Q_{\text{req}} = L_s c_p \int_0^{h_e} \rho(z)[\Theta_E - \Theta(z)] W(z) \; dz
\]

is the energy required to remove the inversion, calculated at sunrise. Here, \(L_s\) is the length of the valley, \(W(z)\) is its width as a function of height, \(h_c\) is the crest height, \(\Theta_E = \Theta(h_c)\), and all other symbols have their usual meanings. The surface sensible heat flux \(H_s\) is defined as normal to a horizontal area, and \(A_p = W(h_c)L_s\). Then, Eq. (7) can be solved to obtain \(t_b\) and the inversion lifetime \(\tau\). Here we assume that \(H_s\) is homogeneous in space, \(\Theta_E\) is constant, and no energy is leaving the valley before breakup—assumptions that are rarely fulfilled.

The condition in Eq. (6) does not hold true since slope winds already export heat before the breakup—a condition that heats the atmosphere above the valley and leads to an increase of \(\Theta_E\) in time. Therefore, the required energy \(Q_{\text{req}}\) [Eq. (8)] is increased by an additional energy \(Q_{\text{add}}\):

\[
Q_{\text{add}} = L_s c_p \int_0^{h_c} \rho(z)[\Theta_E(t_b) - \Theta_E(t_i)] W(z) \; dz,
\]

and the total energy required for the breakup is \(Q_{\text{tot}} = Q_{\text{req}} + Q_{\text{add}}\).

Equation (9) accounts for the energy required to heat a mixed valley atmosphere from \(\Theta_E(t_i)\) to \(\Theta_E(t_b)\). In the case of weak forcing, when the breakup is never reached, the daily maximum of \(\Theta_E\) is taken instead of \(\Theta_E(t_b)\). The necessity of this modification in the case of a nonconstant \(\Theta_E\) was already pointed out by Whiteman and McKee (1982).

Because of the export of heat before breakup, less energy is available to heat the valley, which reduces \(Q_{\text{prov}}\) [Eq. (7)]. In effect, the valley is heated by \(Q_{\text{eff}} = Q_{\text{prov}} - Q_{\text{exp}}\), with

\[
Q_{\text{exp}} = A_p \int_{t_i}^{t_b} [c_p \langle \overline{\theta \theta} \rangle]_{z=h_c} \; dt
\]

being the energy exported out of the valley before the breakup is reached. The horizontal average of the total
vertical heat flux, \([c_F/(\bar{w}\Theta)]_k\), is given in Eq. (5) and is taken at \(z = h_c\). With both \(Q_{\text{add}}\) and \(Q_{\text{exp}}\) taken into account, the breakup is achieved if

\[
Q_{\text{prov}} - Q_{\text{exp}} = Q_{\text{req}} + Q_{\text{add}} \quad \text{at} \quad t = t_b. \tag{11}
\]

The energy exported before the breakup primarily depends on the forcing amplitude \(A_{\text{sw}}\) and heats the atmosphere above the valley. It is important to note that \(Q_{\text{add}}\) and \(Q_{\text{exp}}\) are not the same since \(Q_{\text{exp}}\) describes the export of heat at crest height while \(Q_{\text{add}}\) accounts for additional heat required inside the valley because of a warmer atmosphere above the valley. The initial stratification above the valley will determine the volume that is heated by \(Q_{\text{exp}}\) and therefore the amount of increase in \(\Theta_{\text{e}}\) until breakup. A deep residual layer, which is approximately homogeneously heated by \(Q_{\text{exp}}\), will increase its potential temperature only slightly, in which case \(Q_{\text{add}}\) is small. In Fig. 8, \(Q_{\text{req}}\) and \(Q_{\text{add}}\) are normalized by \(A_p\) and are plotted against \(A_{\text{sw}}\). The initially required energy \(Q_{\text{req}}\) amounts to about 1.6 MJ m\(^{-2}\) times \(A_p\) for all simulations on day 1. After the first night, simulations with a weaker forcing than 500 W m\(^{-2}\) exhibit an increased \(Q_{\text{req}}\) since the elevated stable core has never been removed completely. The cold-air pools that form in the valley for simulations that have reached the breakup on the first day are very similar to each other. Therefore, \(Q_{\text{req}}\) is about equal for these simulations. The additional required energy \(Q_{\text{add}}\), however, depends on the export of heat that increases the temperature above the valley. It rises with \(A_{\text{sw}}\) but reaches a saturation point for simulations that reach the breakup, since the integration time for the calculation of \(Q_{\text{add}}\) becomes shorter. It is up to 1.6 times as large as the initially required energy. On the second day, \(Q_{\text{add}}\) is always smaller than on day 1 and decreases after reaching a maximum at A450 again. This is due to the deep residual layer over the valley that is present in simulations with a strong forcing. On day 2, \(Q_{\text{add}}\) is also much smaller than \(Q_{\text{req}}\) and reaches only 75% of \(Q_{\text{req}}\) for A450. In the case of the A850 simulation, \(Q_{\text{req}}\) on the first day is only about 10% larger than on day 2, whereas the enhancement is approximately sevenfold for \(Q_{\text{add}}\). Since in this simulation the energy exported before breakup is about the same on both days, the differences in \(Q_{\text{add}}\) can largely be attributed to a different stratification above the valley on these two days. A deep residual layer above the valley keeps \(Q_{\text{add}}\) small, which favors an earlier breakup.

Figure 7 shows the predicted inversion lifetime using the thermodynamic bulk model of Whiteman and McKee (1982) without accounting for \(Q_{\text{add}}\) and \(Q_{\text{exp}}\) using Eq. (6) (BU) and with inclusion of the two terms using Eq. (11) (BU-corr). The disagreement between BU and the diagnosed breakup times as compared with BU-corr demonstrates the importance of these two terms. For forcing amplitudes of 450–850 W m\(^{-2}\), neglecting either \(Q_{\text{add}}\) or \(Q_{\text{exp}}\) leads to an underestimation of the inversion lifetime of approximately 0.75–4 h and neglecting both leads to errors of 1.3–4.7 h. The error in the predicted inversion lifetime between BU and BU-corr increases with decreasing forcing amplitude; BU predicts a complete breakup to occur for \(A_{\text{sw}}\) as low as 350 W m\(^{-2}\), whereas the limit is reached in BU-corr and the WRF simulations already at 450 W m\(^{-2}\). The data required for the correction terms in BU-corr come from the WRF Model between \(t_1\) and \(t_b\), and hence BU-corr does not provide a prediction of the breakup time but rather an explanation of why it is reached later than is predicted with BU.

A correct estimate of the breakup time is relevant since the export of heat increases once the breakup is reached. Figure 9 shows the regime diagram and the diurnal evolution of the total vertical heat flux and its components horizontally averaged at 1500 m (crest height) for each simulation. The maximum of the total vertical heat flux appears for most simulations during the MIX and MIXs regime. This is mainly due to the resolved-scale turbulent flux, that is, the transport by convective plumes, which is by far the strongest contributor to the total flux. The mean vertical heat flux—that is, the transport by the cross-valley circulation—is also an important component before the breakup is reached. During the TRA regime, it accounts for about 40% of the total flux. Hence slope winds start to export heat out of the valley up to 2 h before the resolved-scale turbulent flux has increased to a comparable magnitude. The subgrid-scale turbulent flux has a relatively small amplitude and accounts for about 10%–40% of the total.
vertical heat flux. It is most relevant during the transition regime of simulations with a strong forcing.

The fraction of $Q_{\text{prov}}$ that is exported out of the valley atmosphere is shown in Fig. 10 for all forcing amplitudes and the first two simulation days. The provided and exported heat fluxes are hereby integrated before the breakup between $t_i = 0600$ LT and $t_b$, on the basis of Eqs. (7) and (10), respectively. In addition, this fraction is calculated for the period after the breakup by integrating from $t_b$ until $t = 1800$ LT, as well as for the whole day between 0600 and 1800 LT. Note that for simulations that never reach the breakup the integrals over the whole day and before the breakup are identical. Integrated over the whole day, the ratio of exported energy to provided energy increases almost linearly with $A_{sw}$, but more strongly on the second day than on the first day. The ratio for $A_{150}$ on the second day is surprisingly high relative to the first day and to the $A_{200}$ simulation. This result has two reasons: first, katabatic winds, which are stronger in the morning of the second day, import air that is colder than $\Theta_d(z)$ down into the valley, which is counted as a positive heat flux—a contribution that is very small but is nevertheless positive and larger on the second day. Second, the surface sensible heat flux integrated over the whole day is smaller on the second day because of an increased ground heat flux. On day 2, the ratio of $Q_{\text{exp}}$ to $Q_{\text{prov}}$ is much larger for $A_{150}$ when compared with $A_{200}$ since the surface sensible heat flux is negative for several hours for such small forcing amplitudes. Hence $Q_{\text{prov}}$ is very small for $A_{150}$, whereas the katabatic winds and hence $Q_{\text{exp}}$ are of a comparable magnitude for the two simulations.
On the second day, the ratio of exported energy to provided energy lies between 8% and 58% when integrating over the whole day. The ratio after the breakup is about 1.6 times the ratio before breakup. The ratios are much lower on the first day. Because of the almost neutral residual layer on the second day, the stability in the upper third of the valley is higher on the first day, which results in a smaller export of heat.

5. The vertical exchange of mass

Also associated with the inversion breakup is the export of mass and pollutants that may have accumulated in the cold-air pool. In contrast to the total vertical heat flux, which is strongly positive during the day, the horizontally averaged total mass flux at crest height is close to zero since exported mass is replaced by subsiding air from the free atmosphere above the valley. For this reason, we separate the total mass flux

$$\langle \dot{w} \rho \rangle_{\text{TOT}} = \langle \dot{w} \rho \rangle_{\text{ME}} + \langle \dot{w} \rho \rangle_{\text{RES}} + \langle \dot{w} \rho \rangle_{\text{SGS}} $$

into export and import of mass before horizontal integration over two separate areas of positive and negative fluxes:

$$F_{\text{net}}(z = h_c) = \frac{1}{A_p} \int_{A_p^{(+)}} \langle \dot{w} \rho \rangle_{z = h_c} dA$$

$$+ \frac{1}{A_p} \int_{A_p^{(-)}} \langle \dot{w} \rho \rangle_{z = h_c} dA$$

where $\langle \dot{w} \rho \rangle_{z = h_c}$ refers to positive (negative) values of the total vertical mass flux at $z = h_c$ and $A_p^{(+)}$ refers to the part of the total area $A_p$ where positive (negative) vertical mass fluxes occur. Notice that the net vertical mass flux in Eq. (13) is defined per unit area at crest height.

Applying the export and import operators as defined in Eq. (13) to the individual components of the mass flux allows one to quantify the relative importance of each term. The analysis shows that the mean vertical mass flux dominates the budget as the turbulent components are two orders of magnitude smaller. During daytime, the net mean mass flux is positive (out of the valley) but very small, on average approximately $0.2 \times 10^{-3} \text{kg m}^{-2} \text{s}^{-1}$, which is expected since mass is reimported. It is positive, since the valley atmosphere is heated and hence the mean air density decreases, which corresponds to a weak loss of air mass in the valley. This process reverses during the night.

Export and import are very well balanced, but the individual terms can be very large, and maxima of $180 \times 10^{-3} \text{kg m}^{-2} \text{s}^{-1}$ are found in the case of the strongest forcing. The total export of mass is shown with the regime diagram in Fig. 11. Similar to the vertical heat flux, the strongest vertical mass fluxes are found during the MIX and MIXs regimes. Large mass fluxes, approximately one-half as strong as those of the mixed regimes, are also found in the SC regime, and even in the TRA regime a considerable vertical transport is possible. For simulations with a very strong forcing, a mass flux of about $10 \times 10^{-3} \text{kg s}^{-1} \text{m}^{-2}$ is found even during the SEA regime. In these cases, the top of the cold-air pool is located at about 800 m MSL and the nearly neutral residual layer above allows an early development of upslope winds at the upper third of the slope, which implies a positive mass flux. In general, in the case of large upward and downward fluxes, the ventilation of the valley is very good and any pollutants in the valley would be removed very fast. This may not be the case if a very strong but shallow cold-air pool traps pollution close to the bottom of the valley. In this case, mass exchange will happen only above the cold-air pool and will leave the pollution close to the surface unaltered.

To understand the venting of a valley by slope winds, a passive tracer has been added to simulations A150–A850. This tracer is released with an arbitrary but constant emission rate within a box located at the valley floor. This box is approximately 70 m high (seven model levels), is 1.2 km broad in the cross-valley direction (six grid points), and covers the whole valley in the y direction. The emission starts at 0000 LT on the second day to allow the model to accumulate some tracer mass during the buildup phase of the cold-air pool and is kept constant afterward. The tracer concentration is represented by the mixing ratio $\tilde{r}$, and its density shall be

![Figure 11](image-url)
denoted by $\tilde{\rho}_t$. The tracer mass flux through a plane at crest height is equal to the tendency of the total tracer mass above the valley since the only source of tracer mass lies at the valley bottom. Hence

$$F_u(z = h_c) = \frac{1}{m_b} \int_A \left( w\tilde{\rho}_t \right)_{z = h_c} dA = \frac{1}{m_b} \frac{dm_a}{dt},$$  \hspace{1cm} (14)$$

where $m_a$ is the total tracer mass above the valley. The tracer mass flux is normalized by $m_b$, the total amount of tracer mass that is accumulated in the atmosphere from 0000 to 0600 LT. Hence a normalized tracer mass flux of 1 h$^{-1}$ means that the mass $m_a$ is exported out of the valley within 1 h. Figure 12 displays the tracer mass flux, calculated as the tendency of tracer mass above the valley, as a function of time and forcing amplitude. Further, it shows the duration for turning over the whole valley air mass $k$ times.

For simulations with a strong forcing, about one-half of the valley air mass is already exported before the tracer flux increases sharply. This is a result of the deep residual layer, which allows the early development of upslope winds above the cold pool (cf. Fig. 3b) and hence an export of air that contains only small amounts of tracer mass. The net vertical flux of tracer mass is strongest if three conditions are met: upslope winds are strong, large tracer concentrations are available at the valley bottom, and the reinport of tracer mass is prevented. All three conditions are met during the SC regime, which is why the strongest tracer mass fluxes correspond very well to this regime (cf. Figs. 5b and 12). If packets of contaminated air are insufficiently mixed above the valley, a net reinport of tracer mass can occur during the transition from SC to the MIX regime. Once the mixed regime is reached, the tracer mass flux at ridge height drops rapidly since the accumulated tracer has already been exported and the tracer mass flux is now limited by the emission strength at the valley bottom. During the MIX and MIXs regimes, tracer mass released at the valley bottom is constantly mixed upward and any accumulation is inhibited.

The dependence of the vertical exchange of mass on the forcing amplitude is strongly nonlinear. An increase of $A_{sw}$ from 400 to 450 W m$^{-2}$ allows the valley atmosphere to be turned over 2.0 times instead of 1.3 times in 12 h, whereas decreasing the forcing to 350 W m$^{-2}$ results in an export of only 67% of the valley mass during a whole day (Fig. 12). Valleys with weak forcing ($200 < A_{sw} < 400$ W m$^{-2}$) still export some tracer mass as long as the upslope winds are active, but as soon as the normalized export rates drop below the normalized emission rate [1 (6 h)$^{-1}$] tracer mass is accumulated again.

6. Discussion

As long as the atmosphere inside a valley is stably stratified and the synoptic conditions are quiescent, the main exchange mechanism with the free atmosphere aloft is determined by slope winds. The vertical exchange of heat and mass is weaker before the breakup than afterward (cf. Figs. 9 and 11), and hence the breakup time affects the total vertical exchange during the whole day.

The breakup time of a valley inversion primarily depends on the initial stratification and the surface heat budget, as Whiteman and McKee (1982) have shown with a thermodynamic model (see also Fig. 7). The height of the convective boundary layer of the previous day and the depth and strength of the cold-air pool forming during the subsequent night define the initial conditions for the breakup process, which in turn influences the flow patterns emerging during the next day (see section 3b).

Whiteman and McKee (1982) have claimed that their bulk model yields a very good approximation of the inversion lifetime, a conclusion that was confirmed by Bader and McKee (1985). Our simulations agree with this view in general but emphasize the importance of the heat exported by slope winds and the resulting temperature increase above the valley (see section 4). Neglecting these two mechanisms leads to an underestimation of the inversion lifetime of 1.5–4.5 h (see Fig. 7).

The breakup time marks the transition from the SC to the MIX regime, which is the most important for the vertical exchange since vertical fluxes increase once the inversion is removed (cf. Figs. 9 and 11). In general,
the transition of one regime to another is gradual as the valley is heated and the inversion is removed. The regimes SEA, TRA, SC, and MIX are based on the conceptual model by Whiteman (1982) and Bader and McKee (1985). The MIX regime is to our knowledge not yet described in the literature of thermally driven slope flows. It emerges as the convective cells grow horizontally and reach cell diameters of 4–5 km. This behavior is probably related to the growth of the boundary layer, since the cell size of Rayleigh–Benard convection is proportional to the depth of the fluid (Cross and Hohenberg 1993). This regime was not found in simulations with an asymmetric forcing, and we do not expect to find it in an asymmetric valley. It is therefore to some extent an artifact of our idealized model setup. Since the valley atmosphere is completely mixed in the MIX regime, no differences with the MIX regime are found in terms of vertical fluxes.

For a substantial venting of pollutants that have accumulated at the valley bottom during the night, a complete breakup is not required (see Fig. 12), but simulations with a relatively weak forcing, which stay in the SC regime, start to reaccumulate tracer mass after a few hours of decreasing concentrations, since the slope winds are too weak to export tracer mass faster out of the valley than it is generated. The presence of a stable core favors a splitting of the slope flow that leads to a partial recirculation below crest height (Vergeiner and Dreiseitl 1987). The venting of pollutants depends in a nonlinear way on the surface sensible heat flux and therefore on variables that affect the surface heat budget. This mechanism is in good agreement with Lehner and Gohm (2010), who found a strong dependence of the turnover time on the surface albedo and report zero vertical mass fluxes in the presence of a continuous snow cover.

The fact that vertical fluxes of heat and mass increase once the breakup is achieved has implications for errors of models with a mesh size that is too coarse to appropriately resolve the valley. Wagner et al. (2014) have studied the impact of the horizontal resolution on the representation of the valley boundary layer and the evolution of the slope-wind system. For their specific valley geometry, they report that lower horizontal resolutions lead generally to a weaker stability in the valley center. With a 4-km horizontal grid spacing, a mixed regime evolves at a time when an elevated inversion still exists in simulations with a finer grid. An earlier breakup will cause a larger total export of heat, which explains the increased heating rates above crest height that are reported for coarser resolutions. All simulations of Wagner et al. (2014) were performed with identical initial conditions, although the formation of a cold pool during the night is sensitive to the horizontal and vertical resolution as well (Vosper et al. 2013). A weaker cold pool leads to an earlier breakup time and therefore to a stronger vertical heat flux (see Fig. 7 and section 3c). The A150–A850 set of simulations has been rerun with \( \Delta x = 1 \) km using the Mellor–Yamada Nakanishi and Niino level-2.5 boundary layer scheme to verify the results of Wagner et al. (2014) for different forcing amplitudes. The breakup times and amounts of exported energy are similar to our high-resolution simulations since the valley volume is captured fairly well with a 1-km resolution. For most forcing amplitudes, the differences in the breakup time are on the order of 30 min. Together with the findings of our study, the results of Wagner et al. (2014) suggest that the export of heat is overestimated over valleys that are not appropriately resolved and that the timing of the breakup is likely too early. A parameterization of these effects for coarse-resolution models should include a reasonable estimate of the breakup time and reflect the difference of heat and mass fluxes before and after the breakup.

For strong forcing, a considerable amount of the provided energy and mass is exported out of the valley during the day, giving rise to the question of how this affects the heating of the valley atmosphere as compared with the one over a plain. The volume-effect theory or the concept of the topographic amplification factor (TAF; Steinacker 1984) has traditionally been applied in this context. It states that the smaller volume of the valley relative to the plain results in an enhanced diurnal temperature amplitude, assuming that no heat is exchanged with the free atmosphere above. Schmidli (2013) notes that the TAF has to be regarded as the upper limit on the achievable enhancement of the valley temperature amplitude. In his simulation the total heat export was relatively small, amounting to about 15%–20% of the surface heat flux. Hence he concluded that the volume effect is the primary cause of the enhanced diurnal temperature amplitude in mountain valleys. With increasing surface forcing, however, the ratio between provided energy and exported energy increases as well (see Fig. 10), which decreases the contrast between the mean temperature in the valley and over the plain relative to the one diagnosed from the pure volume effect.

In this study, we have chosen a 12-h day, which is only valid at the equinoxes. Keeping the forcing amplitude constant but increasing the day length and hence decreasing the night length would decrease the strength of the cold-air pool in the morning. This would consequently decrease the breakup time and increase the total heat export.

The 2D setup of this study does not allow the development of an along-valley wind. Therefore, the results are not directly applicable at the entrance of a valley.
since the valley wind is strongest at this location and
along-valley advection is supposed to be significant (e.g.,
Schmidli and Rotunno 2010; Wagner et al. 2015b). The
along-valley wind will affect the development of the
cold-air pool during the night and cause cold-air ad-
vection from the surrounding plain during the day,
which will modify the heat balance at a given location.
Nevertheless, we believe that the flow regimes pre-
sented herein develop in the presence of a valley wind as
well but that the timing may differ.

One could expect that asymmetric slope winds due to
an asymmetric forcing could affect the breakup time as
well, but sensitivity runs with a valley that is oriented in
the north–south and east–west directions and is located
at 44°N indicate that this is not the case for the chosen
valley geometry. With a relatively broad valley, heating
by convection from the bottom is dominant and valley
orientation plays no role. This result is in good agree-
ment with Rendón et al. (2015). The development of the
surface inversion has been reported (Zängl et al. 2008;
Vosper et al. 2013) to be dependent on the vertical grid
spacing. Sensitivity experiments in which the height of
the lowest model level is varied between 4 and 12 m
above ground confirm this dependence, which results in
uncertainties of ~1 h in the breakup time.

7. Conclusions

The dependence of the breakup time of a nocturnal
temperature inversion in a valley atmosphere on the
solar forcing is investigated by means of idealized nu-
merical simulations using a horizontal resolution that is
fine enough to resolve the slope-wind system. An ide-
alized solar forcing is applied, and its amplitude is varied
linearly between 150 and 850 W m⁻². The major results
can be summarized as follows:

• Six different flow regimes were identified. The valley
  atmosphere is very stable and hence sealed (SEA) in
  the morning and then goes through a transition regime
  (TRA) until only the stable core (SC) remains inside
  the valley to hinder the vertical exchange. The valley
  atmosphere is eventually mixed (MIX) and may
  reach a variant of this regime with a shifted main-
  updraft zone (MIXs). As the atmosphere is stabilized
  in the late afternoon, a moderately stable (MS) regime
  can be identified.

• The sequence of regimes is mostly the same, regard-
  less of the forcing amplitude, but the strength of the
  forcing determines how long a regime is present in the
  valley atmosphere and which regimes will eventually
  occur. The stratification of the valley atmosphere in
  the morning determines the flow regime after sunrise.

• For the chosen valley topography, the complete
  breakup of the temperature inversion occurs 5 h after
  sunrise for the simulation with the strongest forcing.
  A minimal forcing amplitude of \( \lambda_{sw} = 450 \text{ W m}^{-2} \)
  is required to reach it at all. A deep residual layer above
  the valley favors an earlier breakup. The sensitivity of
  the cold-pool strength on the height of the lowest
  model level causes an uncertainty in the determina-
  tion of the breakup time of approximately ±1 h.

• Averaged over the whole day, between 10% and 57% of
  the energy provided by the surface sensible heat flux
  is exported out of the valley. A larger fraction of the
  provided energy is exported as the forcing is increased.
  For forcing amplitudes of more than 450 W m⁻², the
  ratio of exported energy to provided energy is approx-
  imately 1.6 times as large after the breakup as before it.

• The export of mass strongly increases with increasing
  radiative forcing. Up to 5 times the valley air mass is
  exported during the whole day. If the forcing is strong,
  most pollutants accumulated at the valley bottom are
  removed by upslope winds before the first turnover of
  the valley air has been completed.

Although this study is limited by a simple quasi-2D
topography with one particular height-to-width ratio
and is restricted to dry soil conditions, it gives an over-
view of the sequence of regimes that occur during the
breakup and quantifies the corresponding exchange with
the free atmosphere. We also identified the minimal
energy required for the breakup as well as breakup time.
Furthermore, we highlighted the importance of consid-
ering the heat exported out of the valley before the
breakup is reached when calculating the breakup time.
The exported heat increases the temperature of the at-
mosphere aloft, which results in an additional heating of
the valley atmosphere to reach a mixed state. In a future
study, it would be desirable to investigate the effects of
soil moisture and different initial stratifications on the
formation and destruction of the valley inversion.

Acknowledgments. This work was supported by the
Austrian Science Fund (FWF) under Grant P23918-N21
and by the Austrian Federal Ministry of Science, Re-
search and Economy (BMWFFW) as part of the
UniInfrastrukturprogramm of the Research Platform
Scientific Computing at the University of Innsbruck. We
are also grateful to Simon Vosper for fruitful discussions
at the Met Office. We thank three anonymous reviewers
for their comments that helped to improve this paper.

REFERENCES

Angström, A., 2013: On the counter-radiation of the atmosphere.
Meteor. Z., 22, 761–769, doi:10.1127/0941-2948/2013/0550.
Anquetin, S., C. Guilbaud, and J.-P. Chollet, 1998: The formation and destruction of inversion layers within a deep valley. J. Appl. Meteor., 37, 1547–1560, doi:10.1175/1520-0450(1998)037<1547:TFADIO>2.0.CO;2.

Bader, D. C., and T. B. McKee, 1985: Effects of shear, stability and valley characteristics on the destruction of temperature inversions. J. Climate Appl. Meteor., 24, 822–832, doi:10.1175/1520-0450(1985)024<0822:EOSTAT>2.0.CO;2.

Burns, P., and C. Chemel, 2014: Evolution of cold-air-pooling processes in complex terrain. Bound.-Layer Meteor., 150, 423–447, doi:10.1007/s10546-013-9585-z.

Catalano, F., and C.-H. Moeng, 2010: Large-eddy simulation of the daytime boundary layer in an idealized valley using the Weather Research and Forecasting numerical model. Bound.-Layer Meteor., 137, 49–75, doi:10.1007/s10546-010-9518-8.

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, doi:10.1175/JAM2322.1.

Clements, C. B., C. D. Whiteman, and J. D. Horel, 2003: Cold-air-pool structure and evolution in a mountain basin: Peter Sinks, Utah. J. Appl. Meteor., 42, 752–768, doi:10.1175/1520-0450(2003)042<0752:CASPMT>2.0.CO;2.

Clough, S. A., M. W. Shephard, E. J. Mlawer, J. S. Delamere, M. J. Iacono, K. Cady-Pereira, S. Boukabara, and P. D. Brown, 2005: Atmospheric radiative transfer modeling: A summary of the AER codes. J. Quant. Spectrosc. Radiat. Transfer, 91, 233–244, doi:10.1016/j.jqsrt.2004.05.088.

Colette, A., F. K. Chow, and R. L. Street, 2003: A numerical study of inversion-layer breakup and the effects of topographic shading in idealized valleys. J. Appl. Meteor., 42, 1255–1272, doi:10.1175/1520-0450(2003)042<1255:ANSOIB>2.0.CO;2.

Cross, M. C., and P. C. Hohenberg, 1993: Pattern formation outside of equilibrium. Rev. Mod. Phys., 65, 851–1112, doi:10.1103/RevModPhys.65.851.

Deardorff, J. W., 1980: Stratocumulus-capped mixed layers derived from a three-dimensional model. Bound.-Layer Meteor., 18, 495–527, doi:10.1007/BF00119502.

Dyer, A. J., and B. B. Hicks, 1970: Flux-gradient relationships in the constant flux layer. Quart. J. Roy. Meteor. Soc., 96, 715–721, doi:10.1002/qj.4970964102.

Henne, S., and Coauthors, 2004: Quantification of topographic venting of boundary layer air to the free troposphere. Atmos. Chem. Phys., 4, 497–509, doi:10.5194/acp-4-497-2004.

Hong, S.-Y., and J.-O. J. Lim, 2006: The WRF single-moment 6-class microphysics scheme (WSM6). J. Korean Meteor. Soc., 42, 129–151.

Learae, N. P., E. Crosman, C. D. Whiteman, J. D. Horel, S. W. Hoch, W. O. J. Brown, and T. W. Horst, 2013: The Persistent cold-air pool study. Bull. Amer. Meteor. Soc., 94, 51–63, doi:10.1175/BAMS-D-11-00255.1.

Lehner, M., and A. Gohm, 2010: Idealised simulations of daytime pollution transport in a steep valley and its sensitivity to thermal stratification and surface albedo. Bound.-Layer Meteor., 134, 327–351, doi:10.1007/s10546-009-9442-y.

Niu, G.-Y., and Coauthors, 2011: The community Noah land surface model with multiparameterization options (Noah-MP): 1. Model description and evaluation with local-scale measurements. J. Geophys. Res., 116, D12109, doi:10.1029/2010JD015139.

Noppel, H., and F. Friedler, 2002: Mesoscale heat transport over complex terrain by slope winds—a conceptual model and numerical simulations. Bound.-Layer Meteor., 104, 73–97, doi:10.1023/A:1015556228119.

Paulson, C. A., 1970: The mathematical representation of wind speed and temperature profiles in the unstable atmospheric surface layer. J. Appl. Meteor., 9, 857–861, doi:10.1175/1520-0450(1970)009<0857:TMROWS>2.0.CO;2.

Price, J. D., and Coauthors, 2011: COLPEX: Field and numerical studies over a region of small hills. Bull. Amer. Meteor. Soc., 92, 1636–1650, doi:10.1175/2011BAMS3032.1.

Rendón, A. M., J. F. Salazar, C. A. Palacio, and V. Wirth, 2015: Temperature inversion breakup with impacts on air quality in urban valleys influenced by topographic shading. J. Appl. Meteor. Climatol., 54, 302–321, doi:10.1175/JAMC-D-14-0111.1.

Rotach, M. W., and D. Zardi, 2007: On the boundary-layer structure over highly complex terrain: Key Findings from MAP. Quart. J. Roy. Meteor. Soc., 133, 937–948, doi:10.1002/qj.71.

——, G. Wohlfahrt, A. Hansel, M. Reif, J. Wagner, and A. Gohm, 2014: The world is not flat: Implications for the global carbon balance. Bull. Amer. Meteor. Soc., 95, 1021–1028, doi:10.1175/BAMS-D-13-01091.9.

Schmidli, J., 2013: Daytime heat transfer processes over mountainous terrain. J. Atmos. Sci., 70, 4041–4066, doi:10.1175/JAS-D-13-0831.1.

——, and R. Rotunno, 2010: Mechanisms of along-valley winds and heat exchange over mountainous terrain. J. Atmos. Sci., 67, 3033–3047, doi:10.1175/2010JAS3473.1.

——, and ——, 2012: Influence of the valley surroundings on valley wind dynamics. J. Atmos. Sci., 69, 561–577, doi:10.1175/JAS-D-11-0129.1.

——, and Coauthors, 2011: Intercomparison of mesoscale model simulations of the daytime valley wind system. Mon. Wea. Rev., 139, 1389–1409, doi:10.1175/2010MWR3523.1.

Schnitzhofer, R., and Coauthors, 2009: A multimethodological approach to study the spatial distribution of air pollution in an Alpine valley during wintertime. Atmos. Chem. Phys., 9, 3385–3396, doi:10.5194/acp-9-3385-2009.

Seinfeld, J. H., and S. N. Pandis, 1997: Atmospheric Chemistry and Physics: From Air Pollution to Climate Change. John Wiley and Sons, 1326 pp.

Serafin, S., and D. Zardi, 2010a: Daytime heat transfer processes related to slope flows and turbulent convection in an idealized mountain valley. J. Atmos. Sci., 67, 3739–3756, doi:10.1175/2010JAS3428.1.

——, and ——, 2010b: 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, doi:10.1175/2009JAS3231.1.

Skamarock, W. C., and Coauthors, 2008: A description of the Advanced Research WRF version 3. NCAR Tech. Note NCAR/TN-475+STR, 113 pp. [Available online at http://www.mmm.ucar.edu/wrf/users/docs/arw_v3_bw.pdf.]

Steinacker, R., 1984: Area–height distribution of a valley and its relation to the valley wind. Contrib. Atmos. Phys., 57, 64–71.

——, and Coauthors, 2007: A sinkhole field experiment in the eastern Alps. Bull. Amer. Meteor. Soc., 88, 701–716, doi:10.1175/BAMS-88-5-701.

Vergeiner, I., and E. Dreiseitl, 1987: Valley winds and slope winds observations and elementary thoughts. Meteor. Atmos. Phys., 36 (1–4), 264–286, doi:10.1007/BF01045154.

Vosper, S., E. Carter, H. Lean, A. Lock, P. Clark, and S. Webster, 2013: High resolution modelling of valley cold pools. Atmos. Sci. Lett., 14, 193–199, doi:10.1002/asl.2439.
——, J. K. Hughes, A. P. Lock, P. F. Sheridan, A. N. Ross, B. Jemmett-Smith, and A. R. Brown, 2014: Cold-pool formation in a narrow valley. *Quart. J. Roy. Meteor. Soc.*, **140**, 699–714, doi:10.1002/qj.2160.

Wagner, J. S., A. Gohm, and M. W. Rotach, 2014: The impact of horizontal model grid resolution on the boundary layer structure over an idealized valley. *Mon. Wea. Rev.*, **142**, 3446–3465, doi:10.1175/MWR-D-14-00002.1.

——, ——, and ——, 2015a: Influence of along-valley terrain heterogeneity on exchange processes over idealized valleys. *Atmos. Chem. Phys.*, **15**, 6589–6603, doi:10.5194/acp-15-6589-2015.

——, ——, and ——, 2015b: The impact of valley geometry on daytime thermally driven flows and vertical transport processes. *Quart. J. Roy. Meteor. Soc.*, **141**, 1780–1794, doi:10.1002/qj.2481.

Webb, E. K., 1970: Profile relationships: The log-linear range, and extension to strong stability. *Quart. J. Roy. Meteor. Soc.*, **96**, 67–90, doi:10.1002/qj.49709640708.

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, doi:10.1256/qj.03.214.

——, 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, doi:10.1175/JAM2323.1.

——, ——, and ——, 2007: The effect of mountainous topography on moisture exchange between the surface and the free atmosphere. *Atmospheric Boundary Layers*, A. Baklanov and B. Grisogono, Eds., Springer, 71–88.

Whiteman, C. D., 1982: Breakup of temperature inversions in deep mountain valleys: Part I. Observations. *J. Appl. Meteor.*, **21**, 270–289, doi:10.1175/1520-0450(1982)021<0270:BOTIID>2.0.CO;2.

——, and T. B. McKee, 1982: Breakup of temperature inversions in deep mountain valleys: Part II. Thermodynamic model. *J. Appl. Meteor.*, **21**, 290–302, doi:10.1175/1520-0450(1982)021<0290:BOTIID>2.0.CO;2.

Zängle, G., 2005: Formation of extreme cold-air pools in elevated sinkholes: An idealized numerical process study. *Mon. Wea. Rev.*, **133**, 925–941, doi:10.1175/MWR2895.1.

——, A. Gohm, and F. Obleitner, 2008: The impact of the PBL scheme and the vertical distribution of model layers on simulations of Alpine foehn. *Meteor. Atmos. Phys.*, **99** (1–2), 105–128, doi:10.1007/s00704-007-0276-1.

Zhang, D., and R. A. Anthes, 1982: A high-resolution model of the planetary boundary layer—Sensitivity tests and comparisons with SESAME-79 data. *J. Appl. Meteor.*, **21**, 1594–1609, doi:10.1175/1520-0450(1982)021<1594:AHRMOT>2.0.CO;2.