10.5 A RE-EVALUATION OF THE ROLE OF SUBSIDENCE IN VALLEY AND BASIN WARMING

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1. INTRODUCTION

Subsidence in valleys and basins due to thermally driven upslope flows is usually too weak to be measured directly. Its existence has been established in observational studies mostly through a sequence of vertical temperature profiles showing the downward motion of characteristic features such as inversions. Theoretically, its existence derives from mass-conservation, since air transported upwards by upslope flows on valley sidewalls must be replaced by downward motion in the valley interior. [An alternative way of closing the mass budget would be the horizontal convergence of along-valley flow but observations show this is not the case. During the upslope, up-valley phase there is rather horizontal mass divergence along the main valley, mostly due to the branching out of tributary valleys which all develop up-valley flows (Freytag 1987).] The practical relevance of subsidence is that it can suppress mixed-layer growth in the valley and thereby affect the dispersion of pollutants. In order to quantitatively estimate the strength of subsidence in a given topographic setting one needs an estimate of the upslope mass-flux on the valley sidewalls. An upper limit of this mass-flux can be computed based on the concept of 'equilibrium' slope flow.

2. EQUILIBRIUM SLOPE FLOW

The concept of equilibrium slope flow refers to a balance between diabatic heating and along-slope advection in the slope wind layer

$$\frac{\partial \theta_{SL}}{\partial t} = \frac{H}{c_p \rho D} - U \sin\alpha \frac{\partial \theta}{\partial z} \approx 0 , \qquad (1)$$

where H is the sensible heat flux, D is the slope wind layer depth, U is the upslope flow speed, and α is the slope angle. From (1) the equilibrium upslope mass-flux is given by

$$\rho D U = \frac{H}{c_p} \left(\frac{\partial \theta}{\partial z} \sin\alpha \right)^{-1} . \qquad (2)$$

If such equilibrium is present, it can be shown that the warming of the valley atmosphere due to subsidence corresponds to the sensible heat input at the same height

$$b c_p \rho \left(\frac{\partial \theta}{\partial t} \right)_{SUB} = \frac{H}{\sin\alpha} , \qquad (3)$$

where b is half the width of the valley. In the real atmosphere, slope flows are close to this kind of equilibrium if the slope is steep and homogenous, and the atmospheric stratification is strong. If the slope flow strength is less than the equilibrium value by a factor f, where $0 \le f < 1$, then only a fraction fH of the sensible heat input will re-appear as subsidence warming, and the remaining fraction $(1-f)H$ is available for direct horizontal heat transport towards the valley interior. The resulting total valley warming is again the same as in the equilibrium flow case. Because of this insensitivity of valley warming to the actual slope flow volume flux, simple analytical models (Whiteman and McKee 1982, Haiden 1998) can reproduce the observed temperature evolution in valleys without taking into account the details of the slope flow.

3. COMPARISON WITH METCRAX DATA

Results of a simple numerical model of horizontal heat input and dry-adiabatic adjustment are compared to observations of inversion break-up taken during the Meteor Crater field campaign (METCRAX) conducted in Arizona in 2006 (Whiteman et al. 2008). In valleys the heat budget of a valley cross-section is influenced by along-valley temperature advection. Also, the mass budget of the upslope flow / subsidence circulation is not necessarily closed within a two-dimensional valley cross-section. In closed basins the situation is much more constrained. The upward mass flux across any horizontal plane inside the basin *must* equal the downward mass flux. One has to be careful, however, to identify periods without external influences such as cold air intrusions or intermittent mixing episodes. During the IOP5 morning transition, the temperature evolution did not show disturbances of such kind.

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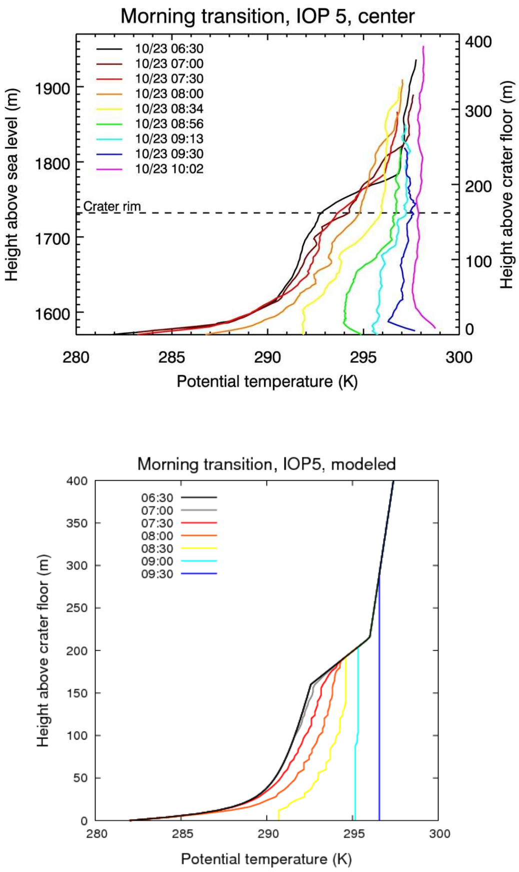

Figure 1. Evolution of potential temperature in and above the meteor crater during the morning transition of IOP5 of METCRAX. Top: observed, bottom: modelled.

As illustrated in Figure 1, the simple horizontal heat-input model with dry-adiabatic adjustment captures the main characteristics of the temperature evolution. In the observations, however, there appears to be a downward movement of the more stable layer near crater rim, especially between 07:00 and 08:56. This process cannot be reproduced by the horizontal heat input method. It also implies a mass-flux that does not vary strongly with height.

Figure 2 illustrates the two limiting cases which bracket the possible set of real-atmosphere conditions. If the upslope mass-flux is in equilibrium (l.h.s. of Figure 2), it becomes smaller as the flow enters a more stable layer and larger as it exits the layer. As a result, detrainment of air from the sidewalls towards the valley interior must take place near the bottom of the stable layer, and entrainment above. Since the flow is in equilibrium, all of the diabatic warming is communicated to the valley interior by the upslope/subsidence circulation, resulting in a quasi-constant warming rate across the stable layer.

In the case of subsidence motion that varies slowly across the stable layer (as suggested by the downward motion of an inversion, r.h.s. of Figure 2), the mass-flux cannot be in equilibrium. Either the local warming caused by the sinking of the stable layer is larger than the diabatic heat input at that height, or the warming in the less stable layers above and below is smaller than the diabatic heat input. In both cases there is no longer a local balance between diabatic heating and vertical advection of background stratification. Below the stable layer, this will lead to a warming of the upslope layer beyond that predicted by the equilibrium solution. Thus, in addition to the background temperature gradient, the gradient of the temperature perturbation needs to be taken into account and (2) be replaced by

$$\rho DU = \frac{H}{c_p}\left(\frac{\partial(\theta+\theta')}{\partial z}\sin\alpha\right)^{-1}. \qquad (4)$$

It can be seen from (4) that in the case of slowly varying surface heat flux and slope angle, a continuous mass-flux across a stable layer can only be achieved by a decrease of perturbation temperature across that layer. Thus, the upslope flow 'uses up' some of its excess temperature to cross the layer. The situation is similar to the well-known mixed-layer growth where near-surface parcels need to attain a higher temperature in order to rise to higher levels.

4. CONCLUSIONS

To a good approximation, the diurnal warming of valleys and basins can be modelled by horizontal heat transport from the sidewalls to the valley (basin) interior. This does not imply that there is in fact a horizontal transfer of air from the sidewalls towards the interior. It is the result of the organized overturning circulation comprised of upslope flow and subsidence motion. It applies to cases where the sidewalls are sufficiently steep so that the upslope mass-flux is close to its equilibrium value. An important practical consequence is that the warming in such cases can be modelled independently of the slope flows. On the other hand, observations which show a downward movement of stable layers in valley and basin atmospheres indicate deviations from equilibrium slope flow. According to (2) this is to be expected at heights where the stratification changes rapidly because the mass-flux cannot adapt to a different value over arbitrarily short vertical distances. If, however, perturbation temperature in the slope flow layer is taken into account, the system has one more degree of freedom, and can adapt to changing stratification by using up (or adding to) the surplus temperature of the slope flow.

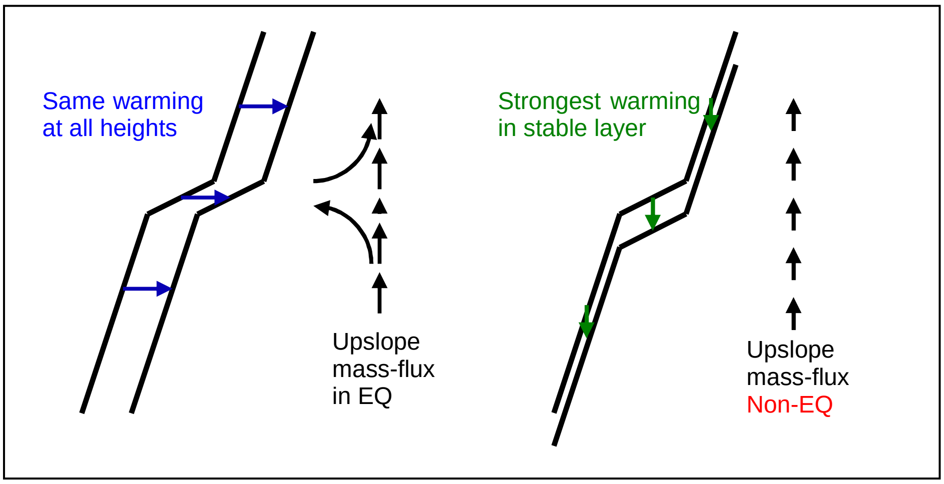

Figure 2. Valley warming in the presence of equilibrium upslope mass-flux (left) and non-equilibrium upslope mass-flux (right). Diabatic heat input at the sidewalls equivalent to a height-independent warming rate has been assumed for simplicity.

Acknowledgements

I would like to thank Dave Whiteman and Sebastian Hoch for providing the METCRAX data and for many inspiring discussions.

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