INVESTIGATING EXCHANGE PROCESSES OVER COMPLEX TOPOGRAPHY

The Innsbruck Box (i-Box)

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In the i-Box project, turbulent exchange processes are investigated experimentally and numerically at their limit, that is, in complex mountainous terrain.

The exchange of mass, momentum, and energy between Earth's surface and the atmosphere occurs through the atmospheric boundary layer (ABL), where the flow is predominantly turbulent. Over flat and horizontally homogeneous terrain turbulent transport is solely responsible for vertical exchange processes, while over areas with significant inhomogeneity or topography, mesoscale flows can contribute to a yet quite unknown degree (Rotach et al. 2014). Moreover, our current theoretical understanding of

turbulent transport processes according to similarity theory (e.g., Holtslag and Nieuwstadt 1986) is largely based on the assumption of horizontally homogeneous and flat (HHF) surfaces. The impact of inhomogeneous surface characteristics as in, for example, land–sea, bare soil–forest, or rural–urban transitions, and the associated internal boundary development, was first studied using a step-change approach (e.g., Garratt 1990) for either a roughness change or in thermal properties (e.g., Gryning and Batchvarova 1990) or a combination thereof (e.g., Savelyev and Taylor 2005). Heterogeneous (patchy) surfaces are often treated using a mosaic (Seth et al. 1994) or tile (Avisar and Pielke 1989) approach in numerical models. Complex surfaces, like those over large roughness elements (vegetation, urban areas), are conceptually being treated by introducing a roughness sublayer below the surface layer (SL) or inertial sublayer (Raupach et al. 1991), wherein the constant flux assumption of Monin–Obukhov similarity theory (MOST; Monin and Obukhov 1954) for the SL does not hold (e.g., Rotach 1993).

As for complex topography, the pioneering approach was to investigate the boundary layer properties over gentle hills (less than 10% slope, say) using a linearization approach (Hunt et al. 1988; Belcher and Hunt 1998). This treatment allows for establishing

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general flow and turbulence characteristics over low hills and has been the basis of a class of relatively simple (mixed spectral finite difference) models (e.g., Mason and Sykes 1979) over gentle topography. For the present purpose, it is important to note that already the treatment of moderate hills introduces an inherent horizontal inhomogeneity in the flow.

Numerical models with a horizontal mesh size down to approximately 1 km, that is, models with parameterized turbulence, employ similarity relations valid for HHF conditions in order to describe exchange processes (e.g., Baklanov et al. 2011), even if the model is applied over strong topography as, for example, for operational numerical weather prediction (NWP) in the Alps, Rockies, or Himalayas. With increasing grid resolution of numerical models, however, the impact of topography becomes more and more relevant, even for models partly (or entirely) resolving turbulence processes. Clearly, when grid resolution increases (mesh

size decreases), topography becomes more realistic and hence steeper so that a number of inherent modeling assumptions may no longer be valid. These include the physical parameterizations (which would, in turn, require physical understanding of the exchange processes over complex topography) as well as the numerical treatment, boundary, and initial conditions and even computational aspects (Arnold et al. 2012).

For any type of topography that exceeds what we may call gentle hills, our current understanding of near-surface exchange processes is not sufficient to honestly evaluate and verify even our high-resolution numerical models. This is true for mesoscale or Reynolds-averaged Navier–Stokes (RANS) equation models as well as partly turbulence-resolving models such as large-eddy simulations (LESSs) or LES-type simulations, that is, with grid spacing somewhere in the gray zone for turbulence [Wyngaard 2004; see Cuxart (2015) for a recent discussion on this issue].

Model evaluation is even more critical if it comes to the verification and possible improvement of what is now called convection-permitting atmospheric models, that is, RANS models with horizontal grid spacing on the order of 1 km, approached by more and more operational centers for numerical weather prediction¹ (e.g., Rotach et al. 2009; Weusthoff et al. 2010). In this case, the model topography is still too coarse to resolve important details (Fig. 1 as an example) but steep enough so that the inconsistencies mentioned above are likely to become relevant. When using observations in complex terrain for model validation—as it is one of the objectives of the present endeavor—extreme

FIG. 1. High-resolution (near real) topography [colors, 100-m-resolution Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) dataset; <https://asterweb.jpl.nasa.gov/data.asp>] as compared to COSMO-I topography (black grid) in the area of the Austrian Inn valley; view from southwest.

¹ At the time of writing this paper operational grid resolution between 1 and 3 km in the horizontal is used—only meteorological services of Alpine countries are reported here—by Deutscher Wetterdienst [model Consortium for Small-Scale Modeling (COSMO; maintains COSMO-Model or COSMO Model; see www.cosmo-model.org), 2.8 km], Italy (different services, model COSMO, 2.8 km), Federal Office of Meteorology and Climatology (MeteoSwiss; COSMO, 1.1 km), Meteo-France [model Application de la Recherche à l'Opérationnel à Meso-Echelle (AROME), 1.3 km], Montenegro [Weather Research and Forecasting (WRF) Model, 1 km], and Zentralanstalt für Meteorologie und Geodynamik [ZAMG (Austrian Weather Service); AROME, 2.5 km].

care must be taken to properly address the problem of different terrain resolutions (the model cannot be expected to reproduce a flow feature that depends on slope angle if the latter is substantially different between real and model topography).

In recent years, a number of projects were performed in order to shed some light on the questions raised above concerning exchange processes of heat, mass, and momentum over truly complex terrain. Among these are the Mesoscale Alpine Programme (MAP) Riviera Project (experimental phase in 1999) in the Alps of southern Switzerland (Rotach et al. 2004), the Vertical Transport and Mixing (VTMX) experimental study in the Salt Lake valley in Utah in 2000 (Doran et al. 2002), the Terrain-Induced Rotor Experiment (T-REX) in the Sierra Nevada (Grubišić et al. 2008), the Convective and Orographically-induced Precipitation Study (COPS) in the Alpine foreland (Wulfmeyer et al. 2011), and, most recently, the Mountain Terrain Atmospheric Modeling and Observations (MATERHORN) Program (Fernando et al. 2015). Additionally, a number of studies with single-site turbulence observations over complex topography are available (e.g., Turnipseed et al. 2003; Martins et al. 2009; Nadeau et al. 2013; Oldroyd et al. 2016). All these have produced extremely valuable datasets through observations and numerical modeling and have led to new insight based on some sample situations [see, e.g., Rotach and Zardi (2007) for a summary of MAP results]. All these efforts, however, lack a systematic, long-term perspective so that findings for a particular flow situation/season/forcing cannot be generalized (or otherwise) and some general conclusions can hardly be drawn.

The Innsbruck Box (i-Box) is an attempt to fill in this gap by establishing a long-term (nonpermanent) experimental setup in complex, mountainous terrain in combination with high-resolution numerical modeling in order to systematically study exchange processes. This combination of experimental activity

FIG. 2. Geographical setting of the i-Box. View from northeast into the Inn valley in western Austria (see its position within the Alps in the inset, time zone is UTC + 1 h). Orange markers: core sites, blue markers: additional sites with turbulence measurements, and green markers: additional sites, mesonet type (see Table 1).

and numerical modeling that is sometimes called the integrative approach is outlined in some detail in the sidebar titled “The i-Box approach.”

The i-Box target area is situated in the Austrian Inn valley and surroundings (see section titled “Experimental layout” and Fig. 2) mainly because of its vicinity to the University of Innsbruck where the project is hosted. Nevertheless, the project’s main focus is not the further investigation of the characteristics of valley (or slope) flows, for which a rich literature and detailed understanding exists [see the excellent textbook by Whiteman (2000) and its follow-up by Zardi and Whiteman (2013)], not least based on studies in the Inn valley (e.g., Vergeiner and Dreiseitl 1987; Zängl 2004; Gohm et al. 2009). Rather, our interest lies in establishing a solid base of knowledge concerning, among others, the following:

- the spatial variability of turbulence characteristics over complex topography (in the horizontal and in the vertical);
- the applicability or possible extension of boundary layer scaling concepts (such as MOST for the near-surface layer);
- the interaction between the local forcing fields (radiation and turbulence) and the resulting (sub) mesoscale flows;

- the relative importance of pure turbulent exchange (as over HHF terrain) and mesoscale flows for the exchange between the complex surface and the free atmosphere (Rotach et al. 2015); and
- the ability of models at different grid spacing in properly reproducing all these phenomena.

Clearly, because of the location of the majority of the i-Box sites and modeling exercises, results will not directly be applicable to other situations of complex terrain such as basins, hilly terrain, mountain ranges, and so on. Still, it is expected that the general features investigated in this project—such as spatial inhomogeneity due to topography, turbulence characteristics over sloped surfaces, and possible extensions to similarity approaches—will be instrumental in gaining insight into the characteristics of these flow situations as well.

The experimental details of the project are explained in the section titled “Experimental layout.” The first scientific highlights—sometimes also encountered problems rather than their solution—and possible applications of the i-Box data pool

are presented in the section titled “First scientific highlights.” Finally, some conclusions and an outlook are presented in the section titled “Outlook.”

EXPERIMENTAL LAYOUT. Recognizing the three-dimensional character of processes to be addressed, i-Box observational sites are located in a 3D box (hence its name) in the Inn valley and surroundings in western Austria. Clearly, assessing the full 3D character of the flow and turbulence state over a box of the present size would exhaust the project’s financial and logistical means. Still, the word *box* in the name emphasizes the need for a full 3D approach and an extension from earlier 1D approaches (e.g., the Lindenberg column; Beyrich and Mengelkamp 2006) for nonmountainous complex surfaces or quasi-2D approaches (as, e.g., in a valley cross section in the Riviera valley; Rotach et al. 2004).

The Inn valley is a major u-shaped valley in the Alps, at the location of the i-Box oriented approximately southwest to northeast, with a width at the valley floor of approximately 2–3 km and a ridge-to-ridge distance of less than 20 km. The valley depth is

THE I-BOX APPROACH.

For meso- β - or meso- γ -scale phenomena in complex terrain, some theoretical understanding is available. For slope flows, for example, the classical theory of Prandtl (1942, 1952) provides a basis for obtaining an analytical formulation for the near-surface wind speed profile, extended over the years to less constrained conditions [most recently, e.g., Grisogono et al. (2015)]. The energetics behind thermally driven flows in valleys or on slopes have been explained using the topographic amplification factor (TAF; Wagner 1932; Steinacker 1984), a concept that has been challenged over the years but recently has been put in context by Schmidli (2013) based on a thorough budgeting approach. He demonstrated that the TAF can be considered an upper limit for the enhanced daily temperature amplitude in a valley, while the associated cross-valley circulation contributes to cooling the bulk of a valley. Given the limitation of the linearized boundary layer theory (Belcher and Hunt 1998) to gentle slopes, however, there is no microscale theory for turbulent

exchange processes in truly complex and steep mountainous terrain.

In the following, TCMT is defined as including (influences of) slopes at different scales steeper than about 10°. Thus, an infinite, homogeneous, and steep slope as the archetypal surface characterization for idealized models [such as the Prandtl (1942) model] will not necessarily respond to the truly complex requirement—even if it may be viewed as the limit for addressing only the first-order (and often leading order) impact of the local slope. We may note that, almost by definition, in TCMT other types of gradients and inhomogeneity (not only slopes) will be present. For the investigation of exchange processes over TCMT, two approaches can be followed: either very high-resolution numerical modeling or measurements.

The numerical modeling approach (Zhong and Chow 2012) has been followed in a number of studies over idealized terrain (e.g., Rampanelli et al. 2004; Catalano and Moeng 2010; Schmidli 2013; Wagner et al. 2014) using horizontal grid spacing of a few hundred meters. For real terrain, however, even

finer resolution would be necessary in order to separate micro- and meso- γ -scale phenomena. Furthermore, even for grid spacing reaching the realms of DNS numerical issues such as the formulation of the surface stress boundary condition (Epifanio 2007) need to be properly addressed and extremely high-resolution surface information is required before results can be considered the truth. Purely experimental investigations, on the other hand, will always suffer from the problem of limited representativeness of observational sites and thus at least render the measurement efforts enormous.

For the i-Box we therefore adopt a combined experimental and modeling approach (Fig. SBI), sometimes referred to as integrated approach. It consists of the following:

- measurements at a number of characteristic sites (see the section titled “Experimental layout” for details) where for a period of several years boundary layer characteristics are monitored in as much detail as possible; and

approximately 1,700–2,100 m, with terrain elevations ranging from about 600 m above mean sea level (MSL) at the valley floor to 2,300 m MSL at mountain ridges to the north and 2,700 m MSL to the south. The valley floor and the lower part of the slopes are characterized by mixed agricultural land intersected by villages of different size and the city of Innsbruck, located at roughly 20 km distance from the majority of sites. The Inn valley is in the transition zone from Mediterranean to continental climate. Daily mean temperature typically ranges from -1°C in winter to $+20^{\circ}\text{C}$ in summer. Mean annual precipitation is about 1,100 mm, 55% of which falls in the convective season (May–September). The period of snow cover in the valley lasts from mid-November to the end of March, and permanent snow cover is typically found from early January to mid-February. The tree line is close to 1,900 m MSL.

Major features in the along-valley direction are the Inn River, the Inntal Highway, and an international rail line. Figure 2 shows the geographical setting of the i-Box sites and their relative arrangement. Locations for individual sites were chosen according to surface characteristics so as to cover characteristic

locations with respect to (mainly) slope, exposition, and land use. No attempt was made to obtain a regular grid of site locations or a profile either across or along the valley direction since this would require a huge number of sites. Furthermore, the chosen experimental design responds to the project goals (i.e., the investigation of exchange processes in complex topography, for which the i-Box sites just provide the most convenient examples).

Sites and instrumentation. We denote as core sites (CS) those that have explicitly been planned and installed for the i-Box. A number of additional sites (AS) either have been added by external partners for particular purposes (such as the Tyrolean government running an air quality warning system for the Inn valley) or are permanent sites in the mesoscale area run by the Institute of Atmospheric and Cryospheric Sciences (ACINN). The core sites are displayed in Fig. 2 and detailed in Fig. ES1 in the supplement to this article (available online at <http://dx.doi.org/10.1175/BAMS-D-15-00246.2>), while an overview of site characteristics is given in Table 1. These site characteristics and

- numerical simulations at very high resolution (VHR) in order to reproduce the full flow characteristics for certain flow situations. The word *reproduce* here ideally means to fall within a predefined error margin at the locations of all sites simultaneously (e.g., Chow et al. 2006) and not only for the mean flow variables but also for the forcing fields such as radiation or turbulence variables [surface fluxes or turbulence kinetic energy (TKE)]. Thus, the goal is to model the right flow for the right reason.

The experimental data alone can be used to investigate turbulence characteristics in complex mountainous terrain [such as Rotach et al. (2008); see also the section titled “First scientific highlights”] as well as overall atmospheric boundary layer characteristics in a mountainous environment (e.g., Weigel and Rotach 2004).

Once the VHR numerical model results are sufficiently accurate those can also be used to obtain data where no measurements are available, thus

complementing the observations. Moreover, the model may yield additional quantities that are difficult to experimentally determine (such as particular terms in the TKE budget equation) or in locations not routinely accessible (e.g., vertical profiles of

turbulence statistics throughout a valley atmosphere). Clearly, any hypothesis arising from analyzing data can be tested using numerical modeling (e.g., Weigel et al. 2007), or sensitivity studies can be performed by systematically varying certain parameters.

FIG. SBI. Conceptual sketch for the i-Box approach. Thin arrows indicate interactions between different types of approaches; thick arrows refer to data flow.

FIG. 3. Photographs of sites (left) CS-NF27 and (right) CS-MT21 (during maintenance). Photo credit: Ivana Stiperski.

their sometimes extreme character—with respect to surface complexity and maintenance and support—are exemplified in Fig. 3 (sites CS-NF27, CS-MT21). The types of measurements available at each site are summarized in Table ES1 in the supplemental information. Finally, Table ES2 lists the used instrument types and their characteristics.

Data from the remote sites (i.e., all sites except AS-VF0; see Table 1) are typically stored on a local datalogger and transmitted to the laboratory several times a day using general packet radio service (GPRS). There turbulence data are stored at the original frequency (20 Hz) and as integral statistics using standard postprocessing. This consists of a fixed averaging interval of 30 min, double-rotation coordinate transformation, and a high-pass filter with a cutoff at 200 s that primarily serves for visual data quality checks. None of the corrections that are usually applied to the integral statistics [flux loss, density, or Webb–Pearman–Leuning (WPL) corrections, etc.; see Aubinet et al. 2012] are applied at this stage. For later convenience, instrument flags are stored with the data. Clearly, for particular applications (e.g., Stiperski and Rotach 2016), postprocessing of turbulence data may have to be performed several times with different postprocessing options.

Data quality is of the utmost importance for any atmospheric measurement and arguably even more so in complex terrain. For all instruments employed within the i-Box, the usual calibrations and factory settings are used as in any experimental setup (Table ES2). In the supplement to this paper, we report on extra efforts that were undertaken in this

respect due to the (type of) instrument that was not previously used in complex terrain or due to using instruments in a long-term setup in rough environmental conditions.

Numerical modeling. Numerical modeling is an important component of the i-Box approach (see Fig. SB1). Numerical models suffer from a number of fundamental and practical problems when used in complex topography (Arnold et al. 2012). While practical problems are tied to computational issues (e.g., numerical stability if slope angles increase), the most fundamental physical problem is the boundary layer (turbulence) parameterization, which most generally is defined based on knowledge from flat and homogeneous conditions. Even the numerical implementation of this knowledge into finite-difference models used in steep terrain requires extreme care (Epifanio 2007), and this slope aspect is most often—to the knowledge of the authors—not taken into account. By definition, in a numerical model a grid average is parameterized as the subgrid-scale turbulence. In extremely high resolution [direct numerical simulation (DNS) with grid spacing of $O(1)$ m], using vertical exchange parameterizations instead of slope normal may be acceptable, but the requirement of surface information at this resolution (in addition to computing requirements) may become an obstacle for any practical application. When very high resolution [LES; grid spacing of $O(100)$ m] is used, slope effects become relevant, but our knowledge on the appropriateness of the common parameterizations (such as MOST) over sloping terrain is limited at best. In addition, even at this horizontal scale, subgrid-scale variability of terrain can have important impacts on the turbulent exchange (e.g., Stiperski and Rotach 2016).

TABLE 1. Basic characteristics of the measurement sites within i-Box. CS refers to core site, AS to additional site, VF is the valley floor, SF is south facing, NF is north facing, MT is mountaintop, and SV is side valley. AfU refers to the Tyrolean environmental protection agency, T stands for turbulence tower (at least one sonic anemometer, at CSs also fast-response hygrometer), R stands for radiation, and M for standard meteorological parameters (wind, temperature, pressure, and humidity). Operational since refers to the date from which measurements were performed; of course, missing data and periods of instrument malfunction may have occurred after that date.

Station name	Type and identification	Local slope (°)	Altitude (m MSL)	Type of observation	Operational since	Surface characteristics
Kolsass	CS-VF0	<0.5	545	T, R, M	Jul 2013	Mixed agricultural
Terfens	CS-SF8	8	575	T, R, M	Sep 2012	Agricultural, car parking
Eggen	CS-SF1	1	829	T, R, M	Apr 2013	Alpine meadow
Weerberg	CS-NF10	10	930	T, R, M	Sep 2012	Alpine meadow
Hochhäuser	CS-NF27	27	1,009	T, R, M	Aug 2013	Alpine meadow
Arbeser	CS-MT21	21 ^a	2,015	T, R, M	Sep 2014	High-alpine vegetation
University ^b	AS-VF0	<0.5	578	T, R, M	HATPRO: Sep 2012 HALO: Jul 2014	Urban (university building)
AfU 1	AS-NF24	24 ^c	867	M	Dec 2013	Alpine meadow
AfU 2	AS-NF23	23 ^d	710	T, M	Dec 2013	Alpine meadow
Ellbögen	AS-SV14	14 ^e	1,069	M	1999	High-alpine vegetation
Sattelberg	AS-SV19	19 ^f	2,111	M	1999	High-alpine vegetation
Obergurglb	AS-SV19b	19	1,938	M	Before 1999	Parking/rural

^a Site on a ridge with indicated slope; steeper slopes to the north (33°) and south (31°) of the site at some distance.

^b The standard meteorological observations (M) are part of ZAMG's Teilautomatische Wetterstationen (TAWES; meaning partly automated weather station) network. Please refer to Table E52 in the supplemental material for more information on the temperature/humidity (HATPRO) and wind (HALO) profiles.

^c Site at the interface of changing slope. Indicated is the slope defined by downslope flow (24°) that feeds into less slanted terrain (11°).

^d Site at the interface of changing slope. Indicated is the slope defined by downslope flow (23°) that feeds into less slanted terrain (16°).

^e Site on a small plateau (local slope negligible). Indicated is the slope defined by nearest downslope flow (14°) at few-meters distance; the outflow of the plateau is steeper (24°).

^f Site on a small plateau (local slope negligible). Influenced by downslope flow from north (16°), northeast (19°), and northwest (24°). The middle value (19°) is referenced in the table.

When it comes to RANS models [grid spacing of $O(1,000)$ m or more] the appropriateness of turbulence parameterizations in truly complex mountainous terrain (TCMT) may even more be questioned than for LES. Generally, the boundary layer and surface exchange treatment in these models is based on the boundary layer approximation, which essentially corresponds to a one-dimensional approach. In RANS models with a TKE turbulence closure, often the conservation equation solved for TKE is one-dimensional (vertical only) so that the impact of heterogeneity (TKE advection, 3D effects) cannot properly be modeled (see the section titled “Real-terrain simulation using COSMO-1” for more details). Also, other physical parameterizations such as that for radiation are usually only 1D for computational

efficiency (Arnold et al. 2012). Finally, semipermanent surface characteristics such as soil moisture (or snow cover) are dependent on terrain characteristics (e.g., runoff, which responds to locally available energy but also to soil type, exposition, ...), and properly taking them into account is by no means trivial but nevertheless can have a crucial impact on model performance (e.g., Chow et al. 2006; Rihani et al. 2015).

While reproducing observed flow characteristics through real-terrain simulations is certainly a first focus when numerical modeling is concerned, simulations using ideal-terrain characteristics—as a form of numerical experimentation—are invaluable for both the experimental planning and data analysis (interpretation). In this sense, Wagner et al. (2015a)

have studied the mean flow and turbulence structure of an idealized 100-km-long and straight valley by systematically varying the height and width in high-resolution (200-m horizontal grid spacing) simulations. The classical (Defant 1949) picture of a single cross-valley circulation cell (slope wind) as it can be found in many textbooks is shown to be prevailing only in relatively shallow or correspondingly broad valleys, while deeper or narrower valleys exhibit two stacked circulation cells separated by an inversion (see also Rotach et al. 2015). While this has been postulated earlier (Vergeiner and Dreiseitl 1987) and observed previously (e.g., Weigel and Rotach 2004; Gohm et al. 2009), the systematic terrain variation allows us to relate the mean flow structure to the corresponding turbulence characteristics and also to the efficiency of heat (energy) exchange between the valley atmosphere and the free troposphere. By introducing spatial variability (the valley narrowing/widening; along-valley slope; Wagner et al. 2015b) or modifying the net available solar energy to investigate the impact of different locations or surface characteristics (Leukauf et al. 2015), these idealized simulations at least partly bridge the gap from the ideal- to the real-terrain setting. Both along-valley geometry variations and slope, as well as the net solar energy input, are shown to have a significant impact on the strength (and reach) of the valley flow and therefore on the overall valley boundary layer structure.

The i-Box data are also used within the COSMO-Next project of MeteoSwiss to evaluate the COSMO model's physics parameterizations (e.g., Buzzi et al. 2011) at the operational grid resolution of 1.1 km. Figure 1 shows that at this resolution at least the major terrain features are resolved, while small side valleys and, especially, mountain peaks are still not appropriately captured. This project has been dubbed Turb-i-Box and consists of choosing a number of characteristic flow states, such as pure valley flow—that is, weak synoptic forcing and development of thermally forced flows within the valley—foehn conditions, situations of pressure-driven channeling (Whiteman and Doran 1993), and nighttime stable boundary layer flows. The long-term observational approach of i-Box allows for choosing several cases per flow type and identifying characteristic model behavior. Thereby, the emphasis does not lie on traditional model verification, in which mean biases or scatter are determined for an average grid point or a question such as “can the model adequately reproduce the valley wind?” is addressed. Rather, the i-Box setting makes it possible to study in detail the possible reasons (contributions) for potential model deficiencies at the small scales, that is, in the driving turbulence, radiation,

and surface fields. Even the contrary, that is, a negligibly small mean bias or scatter, may be found to result from inaccurate boundary layer forcing (compensating errors), thus rendering it at least dangerous for transferability of results to other TCMT sites. It is expected that the results will help to identify generic weaknesses of RANS model turbulence and surface exchange parameterizations in TCMT that later can be reproduced (or otherwise) by other mesoscale models.

Supplementary activities. The layout of the experimental sites (Fig. 2; Table 1) can be considered the backbone of the i-Box, giving rise to a rich dataset over several years of detailed boundary layer observations and results from numerical modeling in TCMT. Despite all efforts, however, such a dataset can never be complete. The vertical dimension of the i-Box in particular suffers from serious shortcomings. Not only would one want to see a continuous monitoring of the valley atmosphere's vertical structure at more than one (AS-VF0) site but also at several along- and cross-valley positions. Even only an occasional scanning of the valley atmosphere using an aerosol or water vapor lidar (e.g., Behrendt et al. 2011) would furthermore be desirable.

The continuous measurement of profiles of integral turbulence statistics from surface-based remote sensing is still in its infancy (Behrendt et al. 2015). Still, a number of recent publications have addressed the issue—most often, however, using full-fledged high-maintenance instruments (e.g., McNicholas and Turner 2014; Behrendt et al. 2015; Wulfmeyer et al. 2010; Harvey et al. 2013). Determining turbulence profiles from commercially available Doppler wind lidars (e.g., the instrument recently installed at site AS-VF0; see Table ES2) is possible in principle but bears large uncertainties (Sathe et al. 2011). Still, for example, Adler and Kalthoff (2014) use profiles of vertical velocity variance (and skewness) to determine the boundary layer height in complex mountainous terrain. An alternative operation mode has recently been proposed by Sathe et al. (2015) and will be tested for mountainous terrain within the i-Box.

The backbone is being supplemented occasionally with additional instrumentation or facilities, be it for educational purposes (e.g., the University of Innsbruck's yearly master of science field course in mountain meteorology; the 2015 Innsbruck Summer School of Alpine Research (InnSAR) summer school on exchange processes in complex terrain] or thematic campaigns. These may, of course, also be conducted by neighboring communities interested in applications over TCMT (hydrology, air pollution modeling, snow modeling, etc.).

A first short campaign of this type was conducted in the fall of 2013 when a number of research flight hours with Deutsches Zentrum für Luft- und Raumfahrt (DLR)'s Cessna Grand Caravan 208B (Mallaun et al. 2015) were realized within and above the Inn valley. During a period of about 2 months around the days scheduled for flight operations, site CS-VF0 was equipped with a sodar (METEK PCS-2000-24/LP) on loan from ZAMG. Furthermore, during the days of flight operations a large aperture scintillometer (Table ES2) was operated across the Inn valley (Fig. ES1). Some preliminary results from this campaign are presented in the section titled “Mean heat flux across a valley” below.

FIRST SCIENTIFIC HIGHLIGHTS. Scaled turbulence statistics on slopes. One of the primary objectives of the i-Box experimental branch is the investigation of turbulence characteristics in TCMT. MOST for the SL cannot a priori be expected to hold due to its assumption of horizontal homogeneity (and constant turbulent fluxes in the vertical), and no corresponding theoretical framework has been put forward so far for complex terrain. Still, the departure from MOST can be studied at the different sites in order to assess whether a similarity approach as such is useful in TCMT and whether additional processes can be identified to supplement the list of key variables (Stull 1988) when seeking a possible extension of MOST for complex terrain. As MOST predicts a universal function for an appropriately scaled mean

variable in the SL, we will treat a failure to follow the HHF curves as a failure of MOST. Still, if the data from a particular site cluster around a well-defined curve, we will call this, following Nadeau et al. (2013), successful local scaling.² It should be noted, however, that identifying statistically significant departures—from HHF curves or between datasets from different sites—is quite demanding with respect to the data quality and availability (Stiperski and Rotach 2016).

Here, we only give an example, that is, present the nondimensional temperature fluctuations $\sigma_\theta/\theta_{*l}$ as a function of (local) stability, that is, z/L_ℓ , where z is the height above ground, L_ℓ is the local Obukhov length, and θ_{*l} the local scaling temperature. For definitions see, for example, Nadeau et al. (2013). A total of 11 months of data that satisfy the criteria for the high-quality dataset of Stiperski and Rotach (2016) are used (i.e., applying all the necessary corrections to the turbulence data: assuring stationarity, obeying a threshold level for statistical uncertainty, and obviously removing periods of hardware failure). Figure 4a shows the results from site CS-NF27 together with a reference curve for HHF conditions (Tillman 1972) and one from Nadeau et al. (2013) based on data over a steep slope. The scatter of the individual data points is considerable (note the double logarithmic representation) but still seems to suggest

² Note that *local* is used here in the sense of Nieuwstadt (1984), where variables at height z are scaled using scaling variables from this height z (and not based on surface fluxes, as in MOST).

FIG. 4. Scaled standard deviation of temperature fluctuations as a function of local stability for unstable stratification from (a) site CS-NF27, data passing the high-quality criterion of Stiperski and Rotach (2016) for an 11-month period between Aug 2013 and Jun 2014 (gray crosses), and (b) from all sites (colored symbols as bin averages, see legend). Error bars refer to the run-to-run variability for each bin (i.e., the standard deviation of all data points falling into this bin). The surface-layer reference (Tillman 1972) and a result for a steep slope (Nadeau et al. 2013) are given for reference and reproduced for the range of applicability as stated by these authors.

some degree of local scaling. In the free convection limit the $-1/3$ slope is indeed observed as predicted by theoretical considerations, not including any assumption about the slope (e.g., Wyngaard 2010). However, the magnitude of negative nondimensional temperature fluctuations appears to be larger than the HHF result (Tillman 1972) and also larger than that from another steep (30° – 41°) slope (Nadeau et al. 2013). For the near-neutral range parameterizations often assumed that a constant value is reached (see Tillman 1972). This, however, would imply that the temperature fluctuations would have to go to zero at the same rate (or faster) as the kinematic heat flux (Tampieri et al. 2009). The present data are at least in line with the latter authors' findings that i) the scatter substantially increases in the near-neutral range and ii) no constant value seems to be approached.

For better visibility, Fig. 4b shows only bin averages (with comparable scatter at sites other than CS-NF27) for five of the i-Box core sites. The results at most of the sites individually and also for the bulk of the dataset (i.e., all sites together) exhibit the $-1/3$ slope for free convection, have generally larger magnitude than the HHF reference, and show a larger scatter (and increasing trend) when approaching neutrality. Site CS-SF8 appears to be special insofar as the free convection slope seems to be violated. The larger values of $-\sigma_\theta/\theta_{*p}$ when compared to HHF values might point to proportionally larger momentum fluxes in steep (mountainous) terrain (note the definition of the temperature scale according to $\theta_{*p} = -\overline{w'\theta'}/u_{*p}$). At all sites, however, the scaled temperature fluctuations appear to be larger when compared to the steep-slope observations of Nadeau et al. (2013). Those are based on a planar fit postprocessing approach, while the present data use the double-rotation option. Stiperski and Rotach (2016) indeed report on systematic differences between the two approaches. Still, what degree this difference (alone) can explain the systematically higher values from the i-Box sites is the subject of a more in-depth investigation to be published elsewhere. As a first tentative conclusion (that will have to be substantiated using more sites and other variables), we may state that scaled temperature fluctuations in TCMT do not conform to MOST, but to some degree local scaling seems to be applicable.

Evolution of temperature structure in a broad valley. The vertical structure of static stability ($\partial\bar{\theta}/\partial z$, $\bar{\theta}$ being the mean potential temperature) is key to the understanding of exchange processes between the valley boundary layer and the free troposphere aloft, be it for daytime conditions as in the idealized

simulations of Wagner et al. (2015a,b) or during the night when a strong surface-based inversion may form, preventing, for example, an effective dispersion of pollutants (typically) emitted near the valley floor (e.g., Gohm et al. 2009). Continuous (operational) observations for model validation or input are still difficult without a highly cost- and personnel-intensive active remote sensing system at hand. Passive remote sensing systems such as a temperature profiler [see the supplement; Humidity and Temperature Profiler (HATPRO) absolute calibration] can bridge this gap, even if they still have difficulty in capturing elevated inversions in particular, which are found from idealized simulations to be characteristic for deep or narrow valleys and to have a profound impact on vertical (heat) exchange (Wagner et al. 2015a). Figure 5 shows the evolution of the temperature structure in the Inn valley for 14–15 December 2012 with a relatively strong surface-based inversion during the night. The ACINN retrieval³ yields a less noisy picture than the original factory retrieval, thus suggesting that the temporal variability in the latter is due to the coarse resolution in the employed (calibration) data rather than real. According to the IMGI retrieval (IMGI is used as an identification of the temperature profiler retrieval), in the afternoon of 14 December a cold pool ($d\theta/dz = 0.02$ – 0.05 K m⁻¹ of about 400-m depth) developed, topped by a shallow near-neutral layer up to about 700 m above ground level (AGL). Higher up and extending far above the ridge top, the stability again decreased ($d\theta/dz \approx 0.005$ K m⁻¹). The factory calibration [radiometer physics GmbH (RPG) retrieval], on the other hand, occasionally even has unstable stratification near the surface during the afternoon of 14 December 2012. After sunset the surface-based inversion started to intensify, the vertical range and intensity of which is verified by the nighttime sounding (0100 UTC 15 December) from the nearby Innsbruck airport (Fig. 5a). The comparison with this sounding nicely demonstrates the improvement in the HATPRO calibration due to employing high-resolution sounding data. It is not so much the accuracy of absolute temperature that is improved (even if it is quite substantially on average; see Massaro et al. 2015) or even the height of the temperature inversion, but it is rather the stability of the layer(s) above the inversion that is largely improved and apparently more realistic when using the improved calibration. The accuracy of information on

³ Note that in Massaro et al. (2015) this retrieval is labeled IMGI (formerly after the Institute of Meteorology and Geophysics, which is now ACINN).

FIG. 5. Evolution of temperature structure within the Inn valley as measured with the HATPRO passive microwave profiler at site AS-VF0 on 14–15 Dec 2012. (a) Potential temperature profiles (line colors: see legend) extracted at the time of the radio sounding from nearby Innsbruck airport; time–height cross sections of potential temperature for (b) RPG retrieval (manufacturer’s calibration; see Table ES2) and (c) IMGI retrieval [improved calibration according to Massaro et al. (2015)]. The bold black lines in (b) and (c) indicate the time of the sounding in (a). The horizontal dashed line marks the mountaintop elevation to the north of the valley. Heights in the figures are AGL.

static stability is crucial when, for example, using the profiler data to assess the ability of numerical models to reproduce the observations (see, e.g., the section titled “Real-terrain simulation using COSMO-1”).

Mean heat flux across a valley. Horizontal inhomogeneity is one of the inherent characteristics of the boundary layer structure in TCMT (Rotach and Zardi 2007). Still, on the valley floor where the thermal conditions often lead to thermally driven along-valley flows, some pseudohorizontal homogeneity (i.e., along the valley axis) may at least be possible above a certain height. During a first period of intensive observations (see the section titled “Supplementary activities”), a scintillometer was operated across the valley over the valley floor site CS-VF0 (Fig. ES1). A scintillometer, the operation principle of which is based on assuming the validity of MOST (e.g., Braam et al. 2012), determines a weighted path-averaged turbulent heat flux, with the largest weight at the center of the path (Pianezze 2013): in this case, approximately the location of the tower based on the valley floor. Over mountainous terrain, Weiss et al. (2001) have shown that (small aperture, in their case) scintillometers can be used

in principle to yield useful data and that even in the SL over slanted surfaces average fluxes of the correct magnitude result (Rotach et al. 2008) as long as the average height of the path is correctly specified. On the other hand, results of Pianezze (2013) from a site with very complex topography (steep, narrow valley) indicate that the obtained structure parameter for temperature can be several orders of magnitude off if the path does not remain within the SL.

At site CS-VF0, even with a flat and relatively homogeneous along-valley footprint, turbulent fluxes are not constant over the height of the tower (i.e., the lowest 17 m) for more than 50% of the time (not shown), thus violating one of the characteristics of the SL. Figure 6 shows the daily cycles of the turbulent heat flux for the three tower levels along with the path-averaged (approximately 1,850-m length) scintillometer-derived turbulent heat flux on two different days. In the early afternoon of 30 August 2013 heat fluxes decrease with height even within the tower layer (Fig. 6a), thus suggesting that at least the 17-m level was already outside the SL. The path-averaged flux as measured at about 60 m AGL by the scintillometer therefore is not within the SL,

and the MOST-based retrieval is likely erroneous. Likewise, momentum flux as expressed by local u_* is also not constant with height but appears to increase above a constant-flux portion near the surface. On 11 September 2013, on the other hand, sensible heat fluxes seem to be constant over most of the day and even the turbulent heat flux from the scintillometer at 60 m AGL is not substantially different (Fig. 6b). Also, within the height range of the tower measurements the momentum fluxes do not show a clear trend with height (but quite some scatter)—at least over the period of the scintillometer measurements (Fig. 6d). Clearly, this result does not allow for a conclusion concerning the horizontal uniformity of turbulent heat flux at the 60-m level; for this, the analysis of the lowest flight legs must be awaited. Still, the results seem to indicate that at least the central part of the near-surface valley atmosphere (where the scintillometer's weighting function is at maximum) can be relatively uniform at times. It remains to be investigated whether a situation with nonconstant fluxes (as on 30 August 2013) simply corresponds to a shallow (possibly local) valley

boundary layer (so that the lowest 60 m cannot be expected to be within the SL) or whether there is another, possibly terrain-related (or slope) reason for those nonconstant fluxes.

Real-terrain simulation using COSMO-I. For a first case study, a period with weak synoptic forcing (pure thermally driven valley flow) was chosen. COSMO-I was run in a configuration similar to the operational setup of MeteoSwiss, that is, with 1.1-km horizontal grid spacing and 80 levels using a terrain-following coordinate system after Leuenberger et al. (2010), and with the lowest model level being located at 20 m. The 40 levels are below 3,000 m (below/within the topography), and the lowest vertical grid spacing is 20 m. The MeteoSwiss preoperational analysis with a horizontal resolution of 1.1 km was used as input data, while the corresponding COSMO-7 forecasts were used as boundary conditions. The domain essentially spans the Alpine range (it is somewhat smaller than the preoperational domain in its southwest corner to save computing power). COSMO has a TKE turbulence closure, that is, a one-dimensional prognostic conservation equation for

FIG. 6. Daily cycles of (a),(b) turbulent heat flux and (c),(d) momentum flux (expressed as local u_*) on (left) 30 Aug and (right) 11 Sep 2013 at site CS-VF0 from point measurements (4, 9, and 17 m AGL) and averaged from a scintillometer across the valley (see green dotted line, Fig. ESI) at 62.4 m AGL (referenced at site CS-VF0).

TKE is solved at half levels (e.g., Buzzi et al. 2011). The physics parameterizations used are the same as in the operational version of MeteoSwiss.

Here, we do not give a full account of the model performance (this will be done elsewhere) but focus only on one particular aspect to emphasize the potential of the i-Box dataset and the various turbulence observations within a small spatial range in TCMT. Figures 7a and 7b depict the TKE time series at the valley floor (CS-VF0) and a steep slope (CS-NF27) site, respectively. The range of TKE values from the nearest grid points (from dark to light shading) is markedly larger on the slope, pointing to the difficulty in choosing an appropriate grid point when comparing with measurements in TCMT. Note that TKE is determined in COSMO at half levels, that is, 10 m in the present setup, and thus corresponds quite closely to the observational heights. While the magnitude of TKE is quite well represented at both sites during the day, the model produces a peak in TKE around the time of flow reversal (from upvalley to downvalley or from upslope to downslope) that is much more pronounced than in the observations (and occurs an hour earlier than at CS-NF27). More

importantly, at both sites (and many of the nearest grid points), modeled TKE steeply falls to very small values during the night, while the observations gradually decrease to moderately small nighttime values around $0.5 \text{ m}^2 \text{ s}^{-2}$. Inspecting the contributions to the TKE budget (Figs. 7c,d) first of all reveals that the

FIG. 7. Time series of TKE at sites (a) CS-VF0 and (c) CS-NF27 on 11–12 Jun 2014 from observations (lines with dots) and COSMO-I simulations. (b),(d) The three terms of the TKE budget equation, that is, shear production (violet), buoyancy (red), and dissipation rate ϵ (orange); see the respective axes (all in $\text{m}^2 \text{ s}^{-3}$). Observed dissipation rate is determined from the spectral densities in the inertial subrange of the velocity spectra. For the simulation results, the full line denotes the nearest grid point, while the dark (light) shaded areas correspond to the 25th (75th)-percentile ranges of the respective variable determined from the nine closest grid points to the observation. The dashed lines depict the median values from those nine nearest grid points. Note that no observed shear term is available at site CS-NF27.

three considered terms (shear production, buoyancy production/damping, and dissipation rate) are astonishingly well captured by the model at both sites and during day and night (at least in this example). Even if one or the other of the budget terms has a slight time shift (e.g., buoyancy production seems to start too late

in the morning of the second day at both sites), none of the budget terms taken into account in COSMO's one-dimensional TKE scheme can be identified as clearly responsible for the too strong decay of TKE after the evening transition. This result would then suggest that a 1D TKE scheme is insufficient in complex mountainous terrain, as has been pointed out previously (Zhong and Chow 2012; Arnold et al. 2012), but not—to the best of our knowledge—demonstrated based on measurements. Clearly, this single-day example will have to be put into context and other situations and flow configurations, and the full range of i-Box sites will have to be taken into account before firm conclusions can be drawn.

Wind input for dispersion modeling in TCMT. ZAMG provides air quality assessment reports for various customers for which wind input information is required. This is a notoriously difficult task in the complex geographical setting of Tyrol. In the framework of an internal ZAMG project, wind fields from different model systems were therefore examined on the basis of i-Box data in order to assess their suitability for dispersion modeling in complex topography. These different model systems included the following:

- Integrated Nowcasting through Comprehensive Analysis (INCA; Haiden et al. 2011) at 1-km spatial resolution and nine levels within the lowest kilometer, which had been developed at ZAMG;
- WRF Model (Grell et al. 2005), also at 1-km spatial resolution and seven levels within the lowest kilometer (no assimilation of any surface data are being used for these runs);
- Eulerian (prognostic) wind fields from the Graz Mesoscale Model (GRAMM; Oettl 2015) at 50-m spatial resolution and 15 levels within the lowest kilometer; and
- the diagnostic fields (enforcing mass consistency) from the Lagrangian Simulation of Aerosol Transport (LASAT; Janicke 2004) as a reference.

All wind fields were interpolated to a horizontal resolution of 50 m.

Here, we focus on the performance at the valley floor and on the slopes of the Inn valley during conditions known to be hard to handle for dispersion modeling, that is, low-wind conditions with uncertain direction of propagation and stronger wind posing numerical challenges in steep terrain. Four case studies, two low-wind periods and two with stronger winds (one connected to a frontal passage, one to a foehn situation), were investigated. LASAT

and GRAMM used data from the i-Box site CS-VF0 (Table 1) at level 2 as input; INCA is nested into the Aire Limitée Adaptation Dynamique Développement International (ALADIN)–AROME model [ALARO; www.rclace.eu/?page=74; ZAMAG's operational weather model] and additionally uses data from the operational observational network (not including i-Box data); the WRF Model is driven from the European Centre for Medium-Range Weather Forecasts (ECMWF)'s global Integrated Forecast System (IFS) model.

One way to obtain quantitative statistical measures for the quality of the wind fields is the comparison with the i-Box stations. For the mentioned case studies, time series of wind speed and direction were derived from all model outputs at the coordinates of the i-Box stations.

Figure 8 shows the near-surface wind field from INCA for the foehn episode (5 February 2014) when southerly flow was channeled into the valleys. Speeds are strongest on the northwest-facing ridges. This corresponds to local experience during foehn conditions (e.g., Gohm et al. 2004) and hence renders the simulated flow field plausible.

An overview of the statistical model evaluation during the frontal passage, the foehn episode, and the two low-wind periods are presented in Fig. 9. Although the sample size is low, some interesting results can be seen. Since the two simpler models LASAT and GRAMM use the wind of station CS-VF0 as input, the uncertainty for the wind direction in the

FIG. 8. Flow field at 10 m AGL (colored arrows corresponding to wind speed, color bar) for the foehn case study [1700–1800 central European time (CET) 5 Feb 2014] as simulated with INCA and interpolated to a horizontal resolution of 50 m. Every eighth wind arrow is shown. Gray solid lines and gray shading correspond to the local terrain height.

valley, that is, at that station but at different levels, is smaller than for the other models. Comparing the two advanced models, INCA shows less bias there. For stronger winds (front, foehn) the simpler models show an opposite bias at the south- and north-facing slopes. Thus, they tend to resemble valley conditions even at the slopes. The winds are quite well captured by INCA for these conditions, except maybe for the north-facing stations during foehn. WRF shows rather high bias and/or root-mean-square error (RMSE) in the valley and on the south-facing slope, while, interestingly, the direction at the two north-facing stations is more accurate. During low-wind conditions, the direction is less defined.

Given the challenging settings used for this study, the results indicate that modeled wind at the valley floor of a major valley like the Inn valley is generally good enough for use in dispersion modeling. The advantage of using a base station within the area to generate the wind field can outweigh the restrictions due to simpler physics. Low wind speeds pose an extra challenge to getting the wind direction right. However, on the slopes the direction of dispersion seems so uncertain that other methods should be taken into account, such as, for example, a set of model realizations with varying wind conditions (i.e., an ensemble approach).

OUTLOOK. The rich i-Box data pool—experimental data and model results and especially their

combination—will give rise to many possible studies (of which quite a few are currently underway). It is expected that the i-Box can contribute to generalizations of earlier findings from episodic studies concerning topics such as the following:

- characterization of the turbulence structure over complex terrain;
- assessment/generalization of some of the classical models for mean flows (slope flows, valley flows, etc.);
- scaling approaches for turbulent exchange on slopes (or generally complex terrain);
- investigation of the relative importance of (fully) turbulent versus so-called submesoscale processes (Mahrt 2014) under (very) stable conditions;
- assessment of model performance (and deficiencies) for mesoscale and even very high-resolution LES numerical models over complex mountainous terrain; and
- provision of benchmark cases (situations) for model intercomparison studies over complex terrain.

So, the i-Box may be called a test bed for the atmospheric boundary layer and related exchange processes over complex mountainous terrain. Clearly, a number of other scientific disciplines will profit from i-Box data and findings as well. Among them are (naturally, as their

FIG. 9. Statistical evaluation for the three types of episodes: (top) frontal passage, (middle) foehn, and (bottom) low-wind conditions. (left) Valley station CS-VF0, (center) south-facing stations CS-SF1 and CS-SF8, and (right) north-facing stations CS-NF10 and CS-NF27. Circles indicate mean bias from the measured wind direction in direction ($^{\circ}$); RMSE in the same unit as error bars. Model results from top: LASAT (L), GRAMM (G), INCA (I), and WRF (W). Gray color indicates that the wind data—though at another level—were taken as input to generate the fields.

interest often is tied to mountainous terrain) catchment hydrology, snowpack modeling (e.g., for avalanche forecast), pollutant dispersion modeling in complex terrain (cf. section titled “Wind input for dispersion modeling in TCMT”), glaciology (glaciers are usually sloped surfaces in mountainous terrain for which classical MOST approaches are employed), statistical or dynamical downscaling approaches (e.g., for climate scenarios) in complex terrain, budgeting approaches for climate applications (e.g., for the global or regional CO₂ budgets; Rotach et al. 2014), and many more.

Apart from future intensive observation activities planned for even more in-depth exploration of the i-Box area over shorter periods using additional instrumental platforms and modeling tools (see the section titled “Supplementary activities”), possible specialized campaigns in one or the other of these neighboring disciplines are very welcome to be conducted within the i-Box facilities. The i-Box may be utilized as providing background information on the distributed boundary layer state over prolonged periods of time, and the interested community would have to add the particular instruments of their interest (for hydrological applications, say, detailed precipitation measurements, runoff, and corresponding soil characteristics, etc.). In any case, the authors will be happy to share i-Box data and information for any application as long as it does not immediately concern one of the institute’s own ongoing Ph.D. theses.

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REFERENCES

- Adler, B., and N. Kalthoff, 2014: Multi-scale transport processes observed in the boundary layer over a mountainous island. *Bound.-Layer Meteor.*, **153**, 515–537, doi:10.1007/s10546-014-9957-8.
- Arnold, D., and Coauthors, 2012: Issues in high-resolution atmospheric modeling in complex topography—The HiRCOT workshop. *Croat. Meteor. J.*, **47**, 3–11.
- Aubinet, M., T. Vesala, and D. Papale, Eds., 2012: *Eddy Covariance: A Practical Guide to Measurements and Data Analysis*. Springer, 438 pp.
- Avisar, R., and R. A. Pielke, 1989: A parameterization of heterogeneous land surfaces for atmospheric models and its impact on regional meteorology. *Mon. Wea. Rev.*, **117**, 2113–2134, doi:10.1175/1520-0493(1989)117<2113:APOHLS>2.0.CO;2.
- Baklanov, A. A., and Coauthors, 2011: The nature, theory, and modeling of atmospheric planetary boundary layers. *Bull. Amer. Meteor. Soc.*, **92**, 123–128, doi:10.1175/2010BAMS2797.1.
- Behrendt, A., and Coauthors, 2011: Observations of convection initiation processes with a suite of state-of-the-art research instruments during COPS IOP8b. *Quart. J. Roy. Meteor. Soc.*, **137**, 81–100, doi:10.1002/qj.758.
- , V. Wulfmeyer, E. Hammann, S. K. Muppa, and S. Pal, 2015: Profiles of second to fourth order moments of turbulent temperature fluctuations in the convective boundary layer: First measurements with rotational Raman lidar. *Atmos. Chem. Phys.*, **15**, 5485–5500, doi:10.5194/acp-15-5485-2015.
- Belcher, S. E., and J. R. C. Hunt, 1998: Turbulent flows over hills and waves. *Annu. Rev. Fluid Mech.*, **30**, 507–538, doi:10.1146/annurev.fluid.30.1.507.
- Beyrich, F., and H. T. Mengelkamp, 2006: Evaporation over a heterogeneous land surface: EVA_GRIPS and the LITFASS-2003 experiment—An overview. *Bound.-Layer Meteor.*, **121**, 5–32, doi:10.1007/s10546-006-9079-z.
- Braam, M., F. C. Bosveld, and A. F. Moene, 2012: On Monin–Obukhov scaling in and above the atmospheric surface layer: The complexities of elevated scintillometer measurements. *Bound.-Layer Meteor.*, **144**, 157–177, doi:10.1007/s10546-012-9716-7.

- Buzzi, M., M. W. Rotach, M. Raschendorfer, and A. A. M. Holtslag, 2011: Evaluation of the COSMO-SC turbulence scheme in a shear-driven stable boundary layer. *Meteor. Z.*, **20**, 335–350, doi:10.1127/0941-2948/2011/0050.
- 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 studies. *J. Appl. Meteor. Climatol.*, **45**, 63–86, doi:10.1175/JAM2322.1.
- Cuxart, J., 2015: When can a high-resolution simulation over complex terrain be called LES? *Front. Earth Sci.*, **3**, 87, doi:10.3389/feart.2015.00087.
- Defant, F., 1949: A theory of slope winds, along with remarks on the theory of mountain winds and valley winds (Zur Theorie der Hangwinde nebst Bemerkungen zur Theorie der Berg-und Talwinde). *Alpine Meteorology: Translations of Classic Contributions by A. Wagner, E. Ekhart, and F. Defant*, C. D. Whiteman and E. Dreiseitl, Eds., Pacific Northwest Laboratory Rep., 95–121. [Available online at www.inscc.utah.edu/~whiteman/homepage/articles/Whiteman1984osti.gov.pdf]
- Doran, J. C., J. D. Fast, and J. Horel, 2002: The VTMX 2000 campaign. *Bull. Amer. Meteor. Soc.*, **83**, 537–551, doi:10.1175/1520-0477(2002)083<0537:TVC>2.3.CO;2.
- Epifanio, C. C., 2007: A method for imposing surface stress and heat flux conditions in finite-difference models with steep terrain. *Mon. Wea. Rev.*, **135**, 906–917, doi:10.1175/MWR3297.1.
- Fernando, H. J. S., and Coauthors, 2015: The MATERHORN: Unraveling the intricacies of mountain weather. *Bull. Amer. Meteor. Soc.*, **96**, 1945–1967, doi:10.1175/BAMS-D-13-00131.1.
- Garratt, J. R., 1990: The internal boundary layer—A review. *Bound.-Layer Meteor.*, **50**, 171–203, doi:10.1007/BF00120524.
- Gohm, A., G. Zängl, and G. J. Mayr, 2004: South foehn in the Wipp valley on 24 October 1999 (MAP IOP 10): Verification of high-resolution numerical simulations with observations. *Mon. Wea. Rev.*, **132**, 78–102, doi:10.1175/1520-0493(2004)132<0078:SFITWV>2.0.CO;2.
- , and Coauthors, 2009: Air pollution transport in an alpine valley: Results from airborne and ground-based observations. *Bound.-Layer Meteor.*, **131**, 441–463, doi:10.1007/s10546-009-9371-9.
- Grell, G. A., S. E. Peckham, R. Schmitz, S. A. McKeen, G. Frost, W. C. Skamarock, and B. Eder, 2005: Fully coupled on-line chemistry within the WRF Model. *Atmos. Environ.*, **39**, 6957–6975, doi:10.1016/j.atmosenv.2005.04.027.
- Grisogono, B., T. Jurlina, Ž. Večenaj, and I. Guttler, 2015: Weakly non-linear Prandtl model for simple slope flows. *Quart. J. Roy. Meteor. Soc.*, **141**, 883–892, doi:10.1002/qj.2406.
- Grubišić, V., and Coauthors, 2008: The Terrain-Induced Rotor Experiment. *Bull. Amer. Meteor. Soc.*, **89**, 1513–1533, doi:10.1175/2008BAMS2487.1.
- Gryning, S.-E., and E. Batchvarova, 1990: Analytical model for the growth of the coastal internal boundary layer during onshore flow. *Quart. J. Roy. Meteor. Soc.*, **116**, 187–203, doi:10.1002/qj.49711649108.
- Haiden, T., A. Kann, C. Wittmann, G. Pistotnik, B. Bica, and C. Gruber, 2011: The Integrated Nowcasting through Comprehensive Analysis (INCA) system and its validation over the eastern Alpine region. *Wea. Forecasting*, **26**, 166–183, doi:10.1175/2010WAF2222451.1.
- Harvey, N. J., R. J. Hogan, and H. F. Dacre, 2013: A method to diagnose boundary-layer type using Doppler lidar. *Quart. J. Roy. Meteor. Soc.*, **139**, 1681–1693, doi:10.1002/qj.2068.
- Holtslag, A. A. M., and F. T. M. Nieuwstadt, 1986: Scaling the atmospheric boundary layer. *Bound.-Layer Meteor.*, **36**, 201–209, doi:10.1007/BF00117468.
- Hunt, J. R. C., S. Leibovich, and K. J. Richards, 1988: Turbulent shear flows over low hills. *Quart. J. Roy. Meteor. Soc.*, **114**, 1435–1470, doi:10.1002/qj.49711448405.
- Janicke, L., 2004: Enhancement of a diagnostic wind field model for licensing industrial facilities (TA Luft). German Federal Environmental Agency UFO-PLAN Project 200 43 256, 196 pp.
- Leuenberger, D., M. Koller, O. Fuhrer, and C. Schar, 2010: A generalization of the SLEVE vertical coordinate. *Mon. Wea. Rev.*, **138**, 3683–3689, doi:10.1175/2010MWR3307.1.
- Leukauf, D., A. Gohm, M. W. Rotach, and J. S. Wagner, 2015: The impact of the temperature inversion breakup on the exchange of heat and mass in an idealized valley: Sensitivity to the radiative forcing. *J. Appl. Meteor. Climatol.*, **54**, 2199–2216, doi:10.1175/JAMC-D-15-0091.1.
- Mahrt, L., 2014: Stably stratified atmospheric boundary layers. *Annu. Rev. Fluid Mech.*, **46**, 23–45, doi:10.1146/annurev-fluid-010313-141354.
- Mallaun, C., A. Giez, and R. Baumann, 2015: Calibration of 3D wind measurements on a single engine research aircraft. *Atmos. Meas. Tech.*, **8**, 3177–3196, doi:10.5194/amt-8-3177-2015.

- Martins, C. A., O. L. L. Moraes, O. C. Acevedo, and G. A. Degrazia, 2009: Turbulence intensity parameters over a very complex terrain. *Bound.-Layer Meteor.*, **133**, 35–45, doi:10.1007/s10546-009-9413-3.
- Mason, P. J., and R. I. Sykes, 1979: Flow over an isolated hill of moderate slope. *Quart. J. Roy. Meteor. Soc.*, **105**, 383–395, doi:10.1002/qj.49710544405.
- Massaro, G., I. Stiperski, I. Pospichal, and M. W. Rotach, 2015: Accuracy of retrieving temperature and humidity profiles by ground-based microwave radiometry in truly complex terrain. *Atmos. Meas. Tech.*, **8**, 3355–3367, doi:10.5194/amt-8-3355-2015.
- McNicholas, C., and D. D. Turner, 2014: Characterizing the convective boundary layer turbulence with a high spectral resolution lidar. *J. Geophys. Res. Atmos.*, **119**, 12 910–12 927, doi:10.1002/2014JD021867.
- Monin, A., and A. M. Obukhov, 1954: Basic laws of turbulent mixing in the surface layer of the atmosphere. Tr. Geofiz. Inst., Akad. Nauk SSSR, 151, 163–187.
- Nadeau, D. F., E. R. Pardyjak, C. W. Higgins, and M. B. Parlange, 2013: Similarity scaling over a steep alpine slope. *Bound.-Layer Meteor.*, **147**, 401–419, doi:10.1007/s10546-012-9787-5.
- Nieuwstadt, F. T. M., 1984: The turbulent structure of the stable nocturnal boundary layer. *J. Atmos. Sci.*, **41**, 2202–2216, doi:10.1175/1520-0469(1984)041<2202:TTSOTS>2.0.CO;2.
- Oettl, D., 2015: Evaluation of the revised Lagrangian particle model GRAL against wind-tunnel and field observations in the presence of obstacles. *Bound.-Layer Meteor.*, **155**, 271–287, doi:10.1007/s10546-014-9993-4.
- Oldroyd, H. J., E. R. Pardyjak, H. Huwald, and M. B. Parlange, 2016: Adapting tilt corrections and the governing flow equations for steep, fully three-dimensional, mountainous terrain. *Bound.-Layer Meteor.*, **159**, 539–565, doi:10.1007/s10546-015-0066-0.
- Pianezze, J., 2013: Modélisation de la structure verticale de la turbulence optique en milieu naturel. Ph.D. thesis, University of Grenoble, 215 pp.
- Prandtl, L., 1942: Führer durch die Strömungslehre. F. Vieweg und Sohn, 382 pp.
- , 1952: Mountain and valley winds in stratified air. *Essentials of Fluid Dynamics*, Blackie and Son, 422–425.
- Rampanelli, G., D. Zardi, and R. Rotunno, 2004: Mechanisms of up-valley winds. *J. Atmos. Sci.*, **61**, 3097–3111, doi:10.1175/JAS-3354.1.
- Raupach, M. R., R. A. Antonia, and S. Rajagopalan, 1991: Rough-wall turbulent boundary layers. *Appl. Mech. Rev.*, **44**, 1–25, doi:10.1115/1.3119492.
- Rihani, J. F., F. K. Chow, and R. M. Maxwell, 2015: Isolating effects of terrain and soil moisture heterogeneity on the atmospheric boundary layer: Idealized simulations to diagnose land-atmosphere feedbacks. *J. Adv. Model. Earth Syst.*, **7**, 915–937, doi:10.1002/2014MS000371.
- Rotach, M. W., 1993: Turbulence close to a rough urban surface part I: Reynolds stress. *Bound.-Layer Meteor.*, **65**, 1–28, doi:10.1007/BF00708816.
- , 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.
- , and Coauthors, 2004: Turbulence structure and exchange processes in an Alpine valley: The Riviera project. *Bull. Amer. Meteor. Soc.*, **85**, 1367–1385, doi:10.1175/BAMS-85-9-1367.
- , M. Andretta, P. Calanca, A. P. Weigel, and A. Weiss, 2008: Turbulence characteristics and exchange mechanisms in highly complex terrain. *Acta Geophys.*, **56**, 194–219.
- , and Coauthors, 2009: MAP D-PHASE: Real-time demonstration of weather forecast quality in the Alpine region. *Bull. Amer. Meteor. Soc.*, **90**, 1321–1336, doi:10.1175/2009BAMS2776.1.
- , 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-13-00109.1.
- , A. Gohm, M. N. Lang, D. Leukauf, I. Stiperski, and J. S. Wagner, 2015: On the vertical exchange of heat, mass and momentum over complex, mountainous terrain. *Front. Earth Sci.*, **3**, 76, doi:10.3389/feart.2015.00076.
- Sathe, A., J. Mann, J. Gottschall, and M. S. Courtney, 2011: Can wind lidars measure turbulence? *J. Atmos. Oceanic Technol.*, **28**, 853–868, doi:10.1175/JTECH-D-10-05004.1.
- , —, N. Vasiljevic, and G. Lea, 2015: A six-beam method to measure turbulence statistics using ground-based wind lidars. *Atmos. Meas. Tech.*, **8**, 729–740, doi:10.5194/amt-8-729-2015.
- Savelyev, S. A., and P. A. Taylor, 2005: Internal boundary layers: I. Height formulae for neutral and diabatic flows. *Bound.-Layer Meteor.*, **115**, 1–25, doi:10.1007/s10546-004-2122-z.
- Schmidli, J., 2013: Daytime heat transfer processes over mountainous terrain. *J. Atmos. Sci.*, **70**, 4041–4066, doi:10.1175/JAS-D-13-083.1.
- Seth, A., F. Giorgi, and R. E. Dickinson, 1994: Simulating fluxes from heterogeneous land surface: Explicit subgrid method employing the Biosphere-Atmosphere Transfer Scheme (BATS). *J. Geophys. Res.*, **99**, 18 651–18 667, doi:10.1029/94JD01330.

- Steinacker, R., 1984: Area–height distribution of a valley and its relation to the valley wind. *Contrib. Atmos. Phys.*, **57**, 64–71.
- Stiperski, I., and M. W. Rotach, 2016: On the measurement of turbulence over complex mountainous topography. *Bound.-Layer Meteor.*, **159**, 97–121, doi:10.1007/s10546-015-0103-z.
- Stull, R. B., 1988: *An Introduction to Boundary Layer Meteorology*. Kluwer, 666 pp.
- Tampieri, F., A. Maurizi, and A. Viola, 2009: An investigation on temperature variance scaling in the atmospheric surface layer. *Bound.-Layer Meteor.*, **132**, 31–42, doi:10.1007/s10546-009-9383-5.
- Tillman, J. E., 1972: The indirect determination of stability, heat and momentum fluxes in the atmospheric boundary layer from simple scalar variables during dry unstable conditions. *J. Appl. Meteor.*, **11**, 783–792, doi:10.1175/1520-0450(1972)011<0783:TIDOSH>2.0.CO;2.
- Turnipseed, A. A., D. E. Anderson, P. D. Blanken, W. M. Baugh, and R. K. Monson, 2003: Airflows and turbulent flux measurements in mountainous terrain: Part 1. Canopy and local effects. *Agric. For. Meteorol.*, **119**, 1–21, doi:10.1016/S0168-1923(03)00136-9.
- Vergeiner, I., and E. Dreiseitl, 1987: Valley winds and slope winds—Observations and elementary thoughts. *Meteor. Atmos. Phys.*, **36**, 264–286, doi:10.1007/BF01045154.
- Wagner, A., 1932: Der taegliche Luftdruck- und Temperaturgang in der freien Atmosphaere und in Gebirgstaelern. *Gerlands Beitr. Geophys.*, **37**, 315–344.
- 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: 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.
- , —, and —, 2015b: 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.
- 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, and M. W. Rotach, 2007: On the nature of turbulent kinetic energy in a steep and narrow Alpine valley. *Bound.-Layer Meteor.*, **123**, 177–199, doi:10.1007/s10546-006-9142-9.
- Weiss, A., M. Hennes, and M. W. Rotach, 2001: Derivation of refractive index and temperature gradients from optical scintillometry for the correction of atmospheric induced problems in highly precise geodetic measurements. *Surv. Geophys.*, **22**, 589–596, doi:10.1023/A:1015644923182.
- Weusthoff, T., F. Ament, M. Arpagaus, and M. W. Rotach, 2010: Assessing the benefits of convection-permitting models by neighborhood verification: Examples from MAP D-PHASE. *Mon. Wea. Rev.*, **138**, 3418–3433, doi:10.1175/2010MWR3380.1.
- Whiteman, C. D., 2000: *Mountain Meteorology: Fundamentals and Applications*. Oxford University Press, 355 pp.
- , and J. C. Doran, 1993: The relationship between overlaying synoptic-scale flow and winds within a valley. *J. Appl. Meteor.*, **32**, 1669–1682, doi:10.1175/1520-0450(1993)032<1669:TRBOSS>2.0.CO;2.
- Wulfmeyer, V., S. Pal, D. D. Turner, and E. Wagner, 2010: Can water vapour Raman lidar resolve profiles of turbulent variables in the convective boundary layer? *Bound.-Layer Meteor.*, **136**, 253–284, doi:10.1007/s10546-010-9494-z.
- , and Coauthors, 2011: The Convective and Orographically-induced Precipitation Study (COPS): The scientific strategy, the field phase, and research highlights. *Quart. J. Roy. Meteor. Soc.*, **137**, 3–30, doi:10.1002/qj.752.
- Wyngaard, J. C., 2004: Toward numerical modeling in the “terra incognita.” *J. Atmos. Sci.*, **61**, 1816–1826, doi:10.1175/1520-0469(2004)061<1816:TNMITT>2.0.CO;2.
- , 2010: *Turbulence in the Atmosphere*. Cambridge University Press, 393 pp.
- Zängl, G., 2004: A reexamination of the valley wind system in the Alpine Inn valley with numerical simulations. *Meteor. Atmos. Phys.*, **87**, 241–256, doi:10.1007/s00703-003-0056-5.
- Zardi, D., and C. D. Whiteman, 2013: Diurnal mountain wind systems. *Mountain Weather Research and Forecasting*, F. K. Chow, S. F. J. DeWekker, and B. Snyder, Eds., Springer, 35–119.
- Zhong, S., and F. K. Chow, 2012: Meso- and fine-scale modeling over complex terrain: Parameterizations and applications. *Mountain Weather Research and Forecasting*, F. K. Chow, S. F. J. DeWekker, and B. J. Snyder, Eds., Springer, 591–653.

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