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Alpine Meteorology

Translations of Classic Contributions by A. Wagner, E. Ekhart and F. Defant

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ALPINE METEOROLOGY: TRANSLATIONS OF CLASSIC CONTRIBUTIONS BY A. WAGNER, E. EKHART, AND F. DEFANT

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ABSTRACT

The English translations of four classic research papers in Alpine meteorology, originally published in German and French in the 1930s and 1940s are presented in this report. The papers include:

- A. Wagner's 1938 paper entitled "Theory and Observation of Periodic Mountain Winds"
- E. Ekhart's 1944 paper entitled "Contributions to Alpine Meteorology"
- E. Ekhart's 1948 paper entitled "On the Thermal Structure of the Mountain Atmosphere"
- F. Defant's 1949 paper entitled "A Theory of Slope Winds, Along with Remarks on the Theory of Mountain Winds and Valley Winds."

A short introduction to these translations summarizes four recent Alpine meteorology field experiments, emphasizing ongoing research that extends the research of Wagner, Ekhart, and Defant. The four experiments include the Innsbruck Slope Wind Experiment of 1978, the MESOKLIP Experiment of 1979, the DISKUS Experiment of 1980, and the ALPEX/MERKUR Experiment of 1982.

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INTRODUCTION

CLASSIC CONTRIBUTIONS TO ALPINE METEOROLOGY

Most of the early work in mountain meteorology was concentrated in the Alps and was published in the German language. The language barrier has made it difficult for the English-speaking meteorologist to access this interesting work, despite the great importance of the early theoretical and observational work which forms the very basis of our present-day understanding of mountain and valley meteorology.

The language barrier has been exacerbated somewhat in the last decade by the de-emphasis of foreign language instruction in university curricula in the United States. For those who have learned to read a foreign language, it is a struggle to keep language skills current when foreign language publications are difficult to find, and one rarely hears a foreign language spoken.

The purpose of this volume is to provide English translations of four of the classical papers on Alpine meteorology originally published in German and French. The papers selected for translation led to the current theory of valley and slope wind systems. They are:

1. Wagner, A. 1938. "Theorie und Beobachtung der periodischen Gebirgswinde [Theory and Observation of Periodic Mountain Winds]", Gerlands Beitr. Geophys., 52, 408-449.

2. Ekhart, E. 1944. "Beitraege zur alpinen Meteorologie [Contributions to Alpine Meteorology]", Meteorologische Zeitschrift, 61(7), 217-231.

3. Ekhart, E. 1948. "De la Structure Thermique de l'Atmosphere dans la Montagne [On the Thermal Structure of the Mountain Atmosphere]", La Météorologie, 4(9), 3-26.

4. Defant, F. 1949. "Zur Theorie der Hangwinde, nebst Bemerkungen zur Theorie der Berg- und Talwinde [A Theory of Slope Winds, Along With Remarks on the Theory of Mountain Winds and Valley Winds]", Archiv fuer Meteorologie Geophysik und Bioklimatologie, Ser. A, 1, 421-450.

Dr. Friedrich Defant has kindly given his permission to reproduce his scientific paper in an English translation. He had previously summarized parts of his paper in an English version in the Compendium of Meteorology under the title of "Local Winds" (Defant, 1951). Dr. Arthur Wagner died in 1942, and Dr. Erwin Ekhart in 1971. The editors of the journals in which their papers first appeared are thanked for permission to publish the present English translations.

Each of the authors of the translated articles was associated with the University of Innsbruck. Dr. Wagner was the Director of the Institute for Cosmic Physics (now the Institute for Meteorology and Geophysics) at the university from 1927 to 1942. Dr. Wagner guided the development of several students whose works have contributed significantly to the development of Alpine meteorology. Dr. Ekhart, an excellent observationalist who has published many research papers (Hoinkes, 1974) in Alpine meteorology, was perhaps the foremost of Dr. Wagner's students. Other students who contributed research in Alpine meteorology include F. Innerebner, A. Jelinek, G. Zuchristian, E. Moll, F. Bondy, A. Burger, E. Schaller, and A. Riedel.

Dr. Defant served as Director of the Weather Service in Innsbruck and as Dozent at the Institute for Meteorology and Geophysics before working with Dr. C. G. Rossby in Stockholm then with Dr. S. Petterssen in Chicago. After 1961 he spent two busy decades as professor at the Institute of Marine Sciences in Kiel, West Germany.

It was perhaps natural that research interests at Innsbruck became focused on Alpine meteorology, since Innsbruck (580 m) is located on the Inn River in one of the major valleys of the Alps. The regular development of the diurnal valley wind systems (Dreiseitl et al., 1980) is a prominent feature of the meteorology of the 2000-m-deep valley, despite the fact that the slope of the valley floor is small. (The Inn Valley falls only 130 m between Innsbruck and the Southern German plains at Rosenheim, a distance of 95 km.) Observational programs were facilitated by the construction of cable cars on both the north- and south-facing slopes of the east-west oriented Inn Valley. The lifts to the Hungerburg, Seegrube, and Hafelekar (2260 m) stations allow access up the south-facing sidewall of the Nordkette Range, while the Patscherkofel lift allows access to a 2250-m peak on the north-facing sidewall. Of course, other aspects of Alpine meteorology have received research attention at Innsbruck, especially the South Foehn, which is especially strong at Innsbruck due to the presence of the Brenner Pass and the Wipp Valley south of Innsbruck.

Wagner's 1938 paper forms the basis of the now accepted theory of along-valley wind systems, explaining their relationship to along-slope wind systems, backing the theory with observations, and explaining the topographic reason for anomalous winds like the ones at Trento, Italy, which had been cited earlier as evidence of a wind system operating in apparent contradiction to his theory.

Ekhart's two papers (1944 and 1948), published after Wagner's death, present experimental evidence which supports Wagner's theories. The 1944 paper deals with aerological observations of valley winds in the Salzach and Lammer valleys of Austria. The 1948 paper presents the results of a series of temperature measurements that allowed, for the first time, the direct testing of Wagner's (1938) theory on the thermal origin of the diurnal along-valley wind systems.

Defant's (1949) paper provides a summary of the knowledge of slope winds. It reviews Wagner's (1938) theory and presents the now well-known set of idealized diagrams illustrating the time and space relationship between the along-slope and along-valley wind systems. He uses composited aerological observations collected by Riedel on the south-facing slope above Innsbruck during both up-slope and down-slope flow periods to show that Prandtl's (1942) steady-state theory of slope currents can explain many of the features of the observations. Defant then extends Prandtl's theory to treat non-stationary conditions.

The papers translated here are only a small sample of a large body of important scientific papers on Alpine metorology. An excellent short summary of much of the early work in Alpine meteorology was published in English in a dissertation written in 1947 at Ohio State University by H. B. Hawkes, entitled "Mountain and Valley Winds With Special Reference to the Diurnal Mountain Winds of the Great Salt Lake Region." Hawkes (1947) gives an especially interesting historical account of the development of present theories of valley and slope wind systems.

RECENT CONTRIBUTIONS TO ALPINE METEOROLOGY

The present volume emphasizes important early work in Alpine meteorology from the 1930s and 1940s. European research in mountain meteorology has, of course, continued up to the present, with some especially interesting contributions to research topics first carefully explored by Wagner, Ekhart, and Defant. The excellent review of mountain weather and climate by Barry (1981) should serve as an introduction to much of the more recent work in these fields. For the interested reader, the proceedings of the biennial European mountain meteorology conferences, held since 1950, provide direct access to recent work. These conference proceedings, up to 1978, are referenced by Barry (1981, p. 16). The two most recent conferences (CIMA80, 1980 and ITAM82, 1982) are included in our list of references. The next conference will be held in Opatija, Yugoslavia in September 1984.

Since 1978, four major international experiments have been conducted in the Alps to investigate the meteorology of Alpine valleys and the development of local wind systems. The valley experiments are:

1. the Innsbruck Slope Wind Experiment of Fall 1978
2. the MESÖKLIP Experiment of Fall 1979
3. the DISKUS Experiment of Summer 1980
4. the ALPEX/MERKUR Experiment of Spring 1982.

These experiments led to theories and research papers that are summarized for experiments 1, 3, and 4 in Freytag and Hennemuth (1983) and for experiment 2 in Fiedler and Prenosil (1980). Brief descriptions of each of these experiments follow.

Innsbruck Slope Wind Experiment 1978

The Innsbruck Slope Wind Experiment (Freytag and Hennemuth, 1979) was conducted from September 29 to October 16, 1978 on the mountainsides and on the floor of the Inn Valley near Innsbruck, Austria. A great deal of earlier work on valley and slope winds had been conducted previously in the same area. Measurements were made in the 1978 experiment using pilot balloons, rawinsondes, a tethered balloon system, instrumented towers, climate stations and motor gliders. A special intensive measurement period was established on October 9 and 10, 1978. The goals of the experiment were:

- to complete the picture of the local circulations under good weather conditions at this location, with special investigation of the slope wind circulations and the role of anti-wind
- to study the growth and thermal structure of the slope wind layer
- to study the evolution with time of the slope and valley wind systems
- to evaluate the influence of temperature inversions on the circulation and the possible development of separate circulations above and below the inversion
- to begin the investigation of valley energetics by evaluating the surface energy budget.

MESOKLIP Experiment 1979

The MESOKLIP experiment (Das Mesoskalige Klimaprogramm im Oberrheintal) was conducted in the upper Rhein Valley from September 17 to 28, 1979. Measurements were made on a single cross section of the valley, located between Karlsruhe and Mannheim, West Germany, at a location where the valley is 35 km wide and the sidewalls are from 100 to 300 m high. Observations were taken on the cross section with multiple pilot balloons launched at 30-minute intervals, with multiple rawinsondes launched at one- or three-hour intervals depending on weather conditions, and with motor gliders making horizontal traverses of the valley. Radiation measurements, surface weather measurements, and measurements from instrumented masts completed the experimental design. Observations were taken on days having one of two types of weather. The first weather type was typical high pressure weather with weak winds and good temperature inversion development. The second weather type had westerlies aloft, winds greater than 3 m/s at a 10-m height in the valley, and weak local thermodynamic influence. The goals of the experiments were:

- to investigate the orographic influences on wind, temperature, and humidity fields
- to investigate the structure in time and space of secondary circulations
- to investigate the buildup and breakdown, as well as the spatial structure, of inversions and low-level jet
- to collect complete data sets for the verification of numerical simulation models.

DISKUS Experiment 1980

From August 6 to 15, 1980, a major cooperative meteorological experiment was conducted in the Dischma Valley of Switzerland. The Dischma Valley is a small, v-shaped, near-linear, inner Alpine side valley which joins the larger valley of the Landwasser River at Davos (Freytag and Hennemuth, 1981, 1982). The Dischma Valley is 14 km long and about 1000 m deep with an inclined valley floor. It runs from SSE toward NNW. The meteorology of the Dischma Valley had been studied previously over a period of years by Urfer-Henneberger (1970) using a network of surface climatological instruments. Urfer-Henneberger developed a detailed conceptual model of the interaction of slope and valley winds and documented the importance of differential solar heating of the two sidewalls on the development of the local wind systems.

The goals of the DISKUS experiment were:

- to construct a three-dimensional scheme for the development of slope winds and mountain and valley winds over time
- to provide details about the energy exchanges at the earth's surface and the coupling of thermal and dynamic effects
- to study the buildup and breakdown of the diurnal regimes.

Data were collected during the DISKUS experiment using pilot balloons, tethersondes, rawinsondes, a SODAR, and motor gliders. A line of fixed climatic measurement stations was located on the valley floor, with two cross valley lines at different distances up the valley. Three energy balance stations were installed in the valley, two on the valley floor and one on the west sidewall. The motor gliders were used during two intensive measurement phases to collect temperature and humidity data above and within the valley. During these measurement phases, the radiating temperature of the valley floor and sidewalls was remotely sensed from an aircraft and from a radiometer located on one sidewall. The goal of the measurement program was to collect wind, temperature, and humidity field data for the whole valley, with the anticipation that the data set from the simple topography of the Dischma Valley would be well suited to the testing of numerical models.

MERKUR Experiment 1982

The MERKUR experiment (Mesoskaliges Experiment im Raum Kufstein/Rosenheim) was a mesoscale experiment conducted as part of the World Meteorological Organization's ALPEX experiment (GARP-ALPEX, 1982). The goals of the MERKUR experiment (Freytag and Hennemuth, 1983) were:

- to study the interactions between the flow in the Inn Valley and the flow in upper Bavaria
- to study the modification of the basic flow by the Alps, especially at the mountain/plain boundary of a major valley
- to investigate thermally induced circulations in the Inn Valley
- to investigate compensation flows between the mountains and plains, especially channeling by the main valleys in the form of anti-wind systems
- to study the structure of the boundary layer in the Alpine foreland.

The MERKUR experiment was conducted in the Inn Valley, in the adjacent mountain areas, and in the Alpine foreland to elevations of 5000 m above sea level. Observations were taken along the length of the valley, from Innsbruck in the interior of the Alps, to Kufstein and Rosenheim, where the Inn flows into the Bavarian foothills. Data were collected within the Special Observing Period of ALPEX from March 23 to April 5. This period included three intensive observation periods of 30 hours each. The intensive observations were conducted during periods of clear weather with well-developed local wind systems and during periods with strong winds aloft. Measuring systems were placed along the Inn Valley from Innsbruck to Rosenheim and along two cross sections, one at the mountain-plain boundary and one higher up in the valley near Rattenberg. In addition, a mountain station that was not influenced by the valley was chosen, and an observational grid was constructed in the foreland.

Measuring systems included surface climate and energy balance stations; upper air sounding stations using pilot balloons, rawinsondes, and tethered balloons; motor gliders and medium-range aircraft; and SODAR and LIDAR sites. Data collected by the various participants of the experiment have been pooled in a data bank established at the University of Munich, West Germany.

REFERENCES

Barry, R. G. 1981. Mountain Weather and Climate. Methuen, London and New York, 308 pp.

CIMA 80. 1980. 16iéme Congrés International de Météorologie Alpine, Aix-les-Bains, September 1980; Contributions proposées (preprints), 462 pp. Société Météorologique de France, 77 rue de Sevres, 92106 Boulogne Billancourt Cedex.

Defant, F. 1951. Local Winds. In Compendium of Meteorology, T. M. Malone (Ed.), American Meteorological Society, Boston, 655-672.

Dreiseitl, E., H. Feichter, H. Pichler, R. Steinacker und I. Vergeiner. 1980. Windregimes an der Gabelung zweier Alpentaler. Arch. Meteor. Geohpys. Bioklima., Ser. B., 28, 257-275.

Fiedler, F. und T. Prenosil. 1980. Das MESOKLIP-Experiment, Mesoskaliges Klimaprogramm im Oberrheintal. Wissenschaftliche Berichte des Meteorologischen Instituts der Universitaet Karlsruhe, Nr. 1, August 1980, 107 pp., Kaiserstrasse 12, D-7500 Karlsruhe, West Germany.

Freytag, C. and B. Hennemuth (Eds.). 1979. Hangwindexperiment Innsbruck, October 1979, Datensammlung. Universitaet Muenchen-Meteorologisches Institut, Wissenschaftliche Mitteilung Nr. 36, 261 pp., Theresienstrasse 37, D-8000 Muenchen, West Germany.

Freytag, C. and B. Hennemuth (Eds.). 1981. DISKUS Gebirgswindexperiment im Dischmatal, Datensammlung, Teil 1: Sondierungen. Universitaet Muenchen, Meteorologisches Institut, Wissenschaftliche Mitteilungen Nr. 43, 250 pp., Theresienstrasse 37, D-8000 Muenchen, West Germany.

Freytag, C. and B. Hennemuth (Eds.). 1982. DISKUS, Gebirgswindexperiment im Dischmatal, Datensammlung Teil 2: Bodennahe Messungen und Flugzeugmessungen. Universitaet Muenchen-Meteorologisches Institut, Wissenschaftliche Mitteilung Nr. 46, 192 pp., Theresienstrasse 37, D-8000 Muenchen, West Germany.

Freytag, C. and B. Hennemuth (Eds.). 1983. MERKUR, Mesoskaliges Experiment im Raum Kufstein - Rosenheim. Universitaet Muenchen-Meteorologisches Institut, Wissenschaftliche Mitteilung Nr. 48, Theresienstrasse 37, D-8000 Muenchen, West Germany, 312 pp.

GARP-ALPEX. 1982. ALPEX Experimental Design, GARP-ALPEX No. 1, World Meteorological Organization, Geneva, 300 pp.

Hawkes, H. B. 1947. Mountain and Valley Winds With Special Reference to the Diurnal Mountain Winds of the Great Salt Lake Region, Ph.D. Dissertation. Ohio State University, 312 pp.

Hoinkes, H. 1974. Erwin Ekhart 1904-1971. Innsbrucker Universitaetsnachrichten IV, Nachrichtenblatt der Universitaet Innsbruck, Studienjahre 1969-1971, Innsbruck, 180-182.

ITAM 82. 1982. 17. Internationale Tagung fuer Alpine Meteorologie Berchtesgaden, Annalen d. Meteorologie, Neue Folge Nr. 19, 293 pp. Offenbach

Prandtl, L. 1942. Stroemungslehre. Verlag Vieweg und Sohn, Braunschweig. 382 pp.

Urfer-Henneberger, C. 1970. Neuere Beobachtungen ueber die Entwicklung des Schoenwetterwindsystems in einem V-foermigen Alpental (Dischma bei Davos). Archiv. Meteor. Geophys. Bioklim., Ser. B, 18, 21-42.

THEORY AND OBSERVATION OF PERIODIC MOUNTAIN WINDS

A. Wagner

Citation:
Wagner, A. 1938. "Theorie und Beobachtung der periodischen Gebirgswinde," Gerlands Beitraege zur Geophysik, 52, 408-449.

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THEORY AND OBSERVATION OF PERIODIC MOUNTAIN WINDS

A. Wagner

Summary

The numerous separate studies of recent years on mountain winds with daily periods are examined to find how well the observed data agree with the theory of valley winds or the direction in which the theory must be expanded or modified. Critical, comprehensive analysis leads to the following results:

1. Valley winds occur in all valleys regardless of their cross-sectional shape. The smaller the moving air mass is, the more these winds are perturbed by local influences. In the case of steep valley floors, a "topographic wind along the sloping valley floor" develops below the true, horizontally flowing valley wind. This reinforces the effect of the lateral slope winds and undoubtedly owes its existence to vertical wind stratification.

2. The influence of valley winds can be sifted out with certainty by averaging, even during overcast weather and during the cold season, and even in the presence of strong gradient winds at the valley floor. A corresponding daily variation of the pressure gradient between plain and valley can also be detected. For the Inn valley, this is still true at the altitude of Hafelekar (2265 m).

3. The theoretical opinion that valley winds represent a closed circulation has been confirmed in all respects: The valley winds appear as local intensifications of a larger circulation system, which during the day cycles air in the lower levels from the surrounding plain toward the mountains, above a neutral region of the mountains and back to the plain again from all points of view. In the same way, every furrow cut into a slope, in which exchange of air with the free atmosphere is weakened, causes reinforcement of slope winds.

4. The region influenced by valley winds generally extends to the crest altitude, although the wind direction may veer below this altitude in the direction of the gradient wind. For high cross winds, the sharp discontinuity between valley wind and gradient wind is probably due to a thermal interface. Such cross winds--even when very weak--result in a lower position of the upper boundary of the valley wind over the middle of the valley than over the slopes.

5. For thermal circulation of valley winds to have an upper limit, a cold zone must overlie the lower warm zone of the valley during daytime and a warm zone must overlie the lower cold zone during nighttime; the thermodynamic origin of this temperature stratification, which is generally required for thermal circulation, will be discussed.

6. Relative to the daily maximum pressure over the plain, the maximum development of the valley wind is delayed, all the more so the less the floor rises, until ultimately the wind maximum of the up-valley wind coincides with the daily pressure minimum in the valley. The phase shift of valley winds with height is minor.

7. Overlapping valley winds, which only blow over the pass altitude into the opposite valley, are more frequent than true Maloja winds, which must flow downwards along the valley floor. Aerologic studies of the Maloja wind in the Oberengadin have shown that it represents a relatively cool current, which veers northeastwards and is superposed by the normal up-valley wind of the Inn valley. The range of the Maloja

wind toward the northeast depends strongly on the general gradient.

8. Valley winds whose origin conforms to the theory of valley winds occur regularly in the circular closed basin of Death Valley.

9. The former scheme given for lateral circulation of slope winds of a valley must be somewhat modified: Air from the valley wind system in the middle of the valley flows into the air mass of the slope winds at all levels in the case of the upslope winds. An average return flow occurs only above the lateral crest height. In the case of the down-valley wind the average return flow occurs only in the lowest cooled layer of air lying on the valley floor. In narrow valleys, therefore, the slope winds extend down to the valley floor, where they influence the daily turning of the wind, whereas in broad valleys the wind roses are not influenced by the slope flows. The phase delay of valley winds relative to slope winds is a necessary consequence of the fact that only circulation of slope winds causes the fairly large temperature fluctuation of valley air.

Since the phase of valley winds is delayed relative to slope winds, the valley winds influence the slopes at the time when the slope winds are shifting. In general, we can state that clockwise circulation of the wind during the day occurs at orographically right-hand slopes and counterclockwise rotation occurs at orographically left-hand slopes.

Wenger's theory is confirmed in the thermal structure of the slope winds: ground-level inversion in the case of downslope wind, strong superadiabatic gradients in the case of up-slope winds. According to the theory, the strongest development of slope winds must be expected at the outer flanks of a mountain range and the weakest in small valleys; the observations to date are therefore in agreement.

10. The abnormal development of valley winds above Trento can be attributed to the influx of relatively cool air from the Sarca valley over a low pass north of Trento; during the day, a cushion of cold air forms in the Etsch valley and permits development of normal up-valley winds at Trento only when this wedge of cold air is compressed northwards.

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The new theory of valley winds [1] formulated in 1932 was at first supported only by results of aerologic studies in the Inn valley [2,3]. A series of individual investigations under the most diverse external conditions was therefore necessary to examine how much the new theory can claim general validity or whether certain regularly occurring features necessitate correction or at least expansion of the theory, which was formulated on the basis of simple assumptions.

It must be emphasized that the numerical estimates [1] on the conditions in the Inn valley are adapted to the size of that valley, and so the resulting conclusions (e.g. influence of slope of valley floor relative to influence of slope of the lateral mountain crests) are valid only for the size conditions of the Inn valley. At that time, the main interest was to demonstrate that strong valley winds could be explained by the theory even in large valleys, where the influence of the almost level valley floor was quite negligible.

These individual studies--particularly by research of the Innsbruck Institute--have now reached a tentative conclusion.

Thus it will be rewarding to subject the earlier theoretical analyses to critical examination in the light of the new and extensive observational data. The purpose of the following lines is not to give a comprehensive review of these research efforts* but only to compare their results with theory. In this connection, we shall also consider some older work in which an interpretation approaching the new theory had already been postulated.

The theory of valley winds derives from the notion, outlined in more detail elsewhere [5], that three genetically quite different phenomena had been considered together up to that time under the term "mountain and valley wind":

1. The shallow slope wind, which flows upwards or downwards along the slopes of isolated mountains and the flanks of a massif, as well as along the lateral slopes of a valley in the line of dip. The old theory of Fournet [6], later rewritten in exact form by R. Wenger [7] using the circulation theorem of V. Bjerknes, was acknowledged as correct for this system. It is worth noting that the shallow winds of the type that frequently occur along the sloping valley floor (during the night, particularly, out of the valley) in short, shallow valleys can be referred to genetically as slope winds along the sloping valley floor, since ultimately any furrow cut into a slope in the direction of dip contributes to reinforcement of the normal slope wind.

It is particularly interesting, however, that such "slope winds along the valley floor" are also found in large valleys, simply if the valley floor has suitable slope. The wind system of the low ground-level layers can be distinguished with certainty from the actual valley wind of the upper layers (see p. 17) by the vertical distribution of wind strength and by the different development phase.

2. The equalizing flow with daily periods, as must develop between lowland plains and flat upland areas due to daily air heating and cooling. Such winds can reach appreciable strength only when the two air masses, which have different vertical thicknesses, are very large relative to the cross section of the horizontal connecting passage. These flows, primarily observed in the high passes of the Himalayas, were first correctly recognized by Sir. R. Strachey [8]. J. v. Hann developed this principle of lifting of surfaces of equal pressure, formulated as early as 1842 by M. Saigey [9], and thus attempted to explain the actual valley winds as well.

3. The valley winds which, particularly in large valleys with almost horizontal valley floor, blow into the valley during the day and out of the valley during the night. A new theory can be formulated for these. It is based on the observation [10] that the air mass in the entire valley cross section experiences greater daily temperature fluctuation than the air at the same height above the plain, and also that the pressure differences developing due to variable heating and cooling during the day largely cancel out above the crest altitude, whereas the pressure gradients effective for valley winds increase in intensity

* Such a review for the layman was recently sent to press by E. Ekhart [4]. Theory and results of field measurements according to the 1935 state of knowledge have already been reported in detail by V. Conrad in Vol. I, Part B of the "Manual of Climatology", edited by W. Koeppen and R. Geiger.

down to the valley floor*. The lateral slope winds are particularly important in this respect, since the heat exchanges at the slope are available to them as energy sources for the necessary vertical air transport.

As early as WWI, pilot balloons had revealed various peculiarities of periodic valley winds. The Villach and Innichen climbs in the Drau valley, which were investigated by L. Barda [12], show that both the up-valley wind and the nighttime down-valley wind reach approximately up to the crest elevation of the lateral mountain ranges. From the same climbs, H. Tollner [13] observed that it is not rare for the transition from up-valley wind to gradient wind to occur via south winds, and interpreted this phenomenon as the "larger valley wind, as wind from the plain perpendicular to the strike of the main Alpine ridge", which is thus in agreement with the views of E. Kleinschmidt [14].

L. W. Pollak [15] and his student V. Wiener [16], who discussed the particularly numerous pilot balloons of Trento for three times of day, obtained results which are difficult to understand. Objectively, they are undoubtedly correct, but they posed a direct obstacle for formulation of the new theory, since these results could neither be reconciled with the existing theories, nor with the new concepts which had gradually been developed. L. W. Pollak [16] wrote: "With increasing altitude, the south wind (up-valley wind) displaces the north wind more and more", and further: "We see therefore that the north winds (down-valley winds) are broken up from above during the day". It was also very apparent from the balloons of Trento that the altitude reached by the valley winds extended considerably above the crest altitude of the adjacent range. In formulating the new theory, the tentative assumption had to be made that a local perturbation was involved here. Proof only became available later (see p. 39).

In 1929 I persuaded Dr. Ebster, at that time Head of the Innsbruck Meteorological Station, to undertake a number of daily releases of pilot balloons with funding made available by the Emergency Association for German Science. At my suggestion, E. Ekhart undertook to analyze these observational data. These results [2], together with some results of supplementary measurements [3], provided the basis for the new theory.

In the following, the individual results of new measurements during recent years will now be compared with the theory.

1. Occurrence of the valley wind in valleys of various sizes and cross sections

Even recently, the predominant view was that valley winds essentially occur in large valleys with U-shaped cross section. Thus, even as late as 1931, H. Tollner [13] wrote: "However, the phenomenon is completely absent...in valleys and canyons which possess steep hillsides but no well-defined valley floor". Thus the theory arrived at the result that the shape of the valley cross section does not in itself influence development of valley winds. Nevertheless, Tollner also expressed the

* I would like to use the opportunity to mention a 1932 article of E. v. Everdingens [11], which I became aware of only after completing my theoretical investigations. The author emphasizes there that, as early as 1914, he had expressed the view that the driving force of valley winds consists in stronger heating of the valley air in the entire valley cross section than of the air above the plain, and that horizontal pressure gradients thus develop between plain and valley, increasing from crest altitude down to the valley floor. These notions coincide perfectly with the basis on which the new theory is developed.

suspicion that the valley wind might be suppressed in steeply ascending, mostly V-shaped valleys, if the lateral slope winds, which indeed can have only limited layer thickness, no longer transport sufficient air vertically upwards or downwards.

Subsequent observations [17] have shown that the shape of the valley cross section actually is of no importance: in the Ötz valley, valley winds were observed just as surely in the upper V-shaped part as in the lower U-shaped part. In steeper valleys [18], however, as in the Kauner and Watten valleys, in the short Vikar valley (only 9 km long) and in the Hall valley, which is also short, the valley wind occurs as an absolute fact of nature, although it is not as regular as in large valleys with almost horizontal valley floors. The following relationship became apparent between valley length and size of valley cross section: The larger the air mass set in motion, the smaller the influence of local perturbations limited to a small area. In large, long valleys, as in the Inn valley, the valley winds develop extremely regularly--the measurements of a single good summer day are good enough to provide an absolutely typical picture--whereas in short valleys, such as the Vikar valley, even the shadow of a fairly large cloud can greatly weaken or completely arrest the flow. Moreover, the wind flowing in the valley direction in the innermost part of a valley has to break up into individual slope wind circulations which vary greatly in time.

E. Ekhart [19] has given the obvious explanation for the occurrence of valley winds despite steeply sloping valley floor: The sloping valley floor gives rise to shallow, ground-level flow in precisely the same way as the lateral slopes, because insolation during the day and ground radiation during the night release those energies at such valley floors which are necessary for vertical transport of the adjacent air. In fairly steep valleys, therefore, this shallow "slope wind along the valley floor" (as Ekhart calls it), reinforces the effect of lateral slope winds.

Interestingly, these shallow "slope winds along the valley floor" can sometimes be distinguished quite clearly from the actual valley wind blowing above them: In common with slope winds on the whole, this shallow flow along the sloping valley floor also starts much earlier in the morning and evening than the actual valley winds, so that--for example at Laengenfeld in the Ötz valley [17]--a shallow down-valley wind was observed in the evening in the lowermost layers, although the up-valley wind above those levels was still fully developed.

It can be expected that this "slope wind along the valley floor" also differs in thermal structure from the valley wind blowing above it: The former generally ought not to reach any higher than the nighttime ground-level inversion or, during daytime, than the layers with superadiabatic temperature gradients. During the night, this interface can be perceptible as a distinct wind minimum due to slight friction between the two flows, whereas during the day, it is generally obscured by vigorous vertical mixing.

This wind minimum has also been found in the up-valley wind, but only in fairly steeply sloping valley floors. This division into two layers appears very instructively in the graphical diagram for Stuben, on the western slope of Arlberg pass (see Figure 2 in [19]). Characteristically, this Figure shows that the maximum wind strength occurs an hour earlier in the ground-level layer than in the main valley wind layer. The wind minimum occurs at that altitude at which the strength of the actual valley wind increases with altitude due to decreasing ground friction by just as much as the strength of the shallow "slope wind along the valley floor" decreases with altitude.

Judging from existing experience in various Alpine valleys, we can expect that valley winds--apart from the low nighttime ground-level flow--would also be found up to approximately the lateral crest altitude even in small valleys, such as those of the German Mittelgebirge (foothills) (e.g. Breusch valley or Wisper valley) by careful aerologic investigation. In the case of fairly strong gradient winds, there would at least be a difference effect between afternoon and morning. The fact that these local winds are completely obscured in individual balloon soundings and that up-valley winds in particular are not perceptible even in the vicinity of the valley floor ought to be due to the following reason: According to the theory, the effective thermal pressure differences are smaller the lower the relative crest altitude is. Consequently, even slight gradients in overall weather conditions can suppress development of such valley winds. This is particularly true during daytime, when vertical exchange is intensive due to rapid warming. Penetration of the general gradient wind into such low valleys will naturally occur more easily if the valley direction approximately coincides with the direction of the mean gradient wind, as is almost the case in both the Breusch [20] and Wisper [21] valleys. Nevertheless, an up-valley wind does develop in the Wisper valley even under clear skies (H. Schultz [21]), although as the author emphasizes, it is "completely unnoticed".

We can now assert that a valley wind system can be observed in all valleys in which such investigations have been initiated on the basis of aerologic measurements or hourly values of the ground-level wind. This has been proven by the studies of A. Jelinek [22] in the Southern Tyrol, by the aerologic measurements in the Gail valley [23], in the Schwarza valley [24,25,26] and in the Rausier and Moell valley north and south of Sonnblick [27], apart from the already mentioned studies in other valleys.

To obtain proof that all the various features of vertical structure and daytime development of valley winds observed in various Alpine valleys are also valid outside the Alps, I requested my colleague Kinzl to release a series of pilot balloons in interconnected series of valleys during his 1936 Andes expedition [28] in the Rio Santa valley between Cordillera Blanca and Cordillera Negra. His observations fully confirm the results obtained in the Alps. The analysis of these balloon releases is largely complete, although the results have not yet been published.

2. Dependence on season and weather

Until recently, valley winds have generally been interpreted as a phenomenon of good summer days. Certainly they develop most strongly during clear, warm weather. However, if the theory is correct in the sense that the pressure gradients are due to different daily temperature fluctuation in the valley air and in the air above the plain, then at least traces of valley winds should be perceptible even in the cold season and even during overcast weather. This is in fact the case: Analyses of wind stations in the Southern Tyrol [24] yield wind figures for two stations which, even for overcast weather, absolutely suggest a regular valley wind. Moreover, H. Tollner [13] had shown as early as 1931 that valley winds can occur throughout the year at Salzburg, except for December and January.

For the Southern Tyrol, A. Defant [29] had shown as early as 1909 not only that the valley winds there are unmistakable even in winter, but also that the corresponding daily variation of pressure gradients is observed in winter. Of course, the insolation conditions there are much more favorable than north of the central Alpine crest. For Innsbruck, the dependence of valley winds on season and weather has been thoroughly investigated by F. Bondy [30]. The ratio of east wind occurrences to the west wind occurrences here around 1400 hours is considerably greater than during the morning in all calendar months. Moreover, there is a daily variation of the Munich-Innsbruck pressure difference, suggesting that weak development of valley winds can be expected even in winter.

Admittedly, the daily average pressure gradient in winter causes winds flowing out of the valley in the Innsbruck to Munich direction, whereas the average gradient in summer has the opposite direction. In winter, therefore, the Munich-Innsbruck pressure difference is positive only for a few hours of the day, whereas during the night the not inconsiderable Innsbruck-Munich pressure difference at ground level only occasionally--because of the great friction and stable air stratification--results in corresponding air movement. At higher altitudes, however, the valley winds exist quite unmistakably throughout the entire year, as an investigation of E. Ekhart [31, Fig. 7] demonstrates.

Bondy described very interesting indirect observations of the valley wind in the presence of strong gradient winds, which prevent a wind shift during the day: On days when the foehn blows without interruption along the valley floor, the influence of the periodic valley wind is manifested by a distinct leftwards shift of wind direction between night (zero to 600 hours) and day (1200 to 1800 hours). The shift is around 12° in summer and somewhat less in winter. Thus a small west component (down-valley wind) is superposed on the gradient wind from the south during the night, and a small east component (up-valley wind) during the day.

Pilot balloons of R. Kanitscheider [32,33] on foehn days have already revealed that the west wind, which at Innsbruck blows in foehn weather not only before the foehn descends to the valley floor, but also during the nighttime foehn pauses, is essentially the nighttime down-valley wind of the upper Inn valley. When the temperatures here have dropped sufficiently far below those of the foehn flow due to ground radiation, this wind streams in below the foehn and pushes it to higher altitudes. This west wind only has significant strength at Innsbruck where its streamlines strongly converge in the vertical because of the deeply penetrating foehn. These foehn pauses with west wind generally begin in the late evening hours, and frequently only after midnight, and last until around noon. Admittedly, the phenomenon is not always a pure valley wind, since this west wind occasionally persists for the entire day. Thus it is conceivable that the general pressure distribution during foehn weather causes outflow of air from the northern Alps towards the Bavarian high plain. However, the fact that these west winds are most frequent and strongest in the morning hours can probably be attributed to the daily variation of the pressure gradients which produce the valley winds. Of course, the up-valley wind cannot be directly observed on foehn days, because it mixes with the divergent foehn flow.

The influence of valley winds is manifested in similar fashion in the daily variation of wind strength on days with continuous west wind on the one hand and on days with east winds on the other: Whereas the daily maximum of the west winds occurs as early as 1300 hours, that of the east winds only occurs at 1600 hours. When both groups of days are combined, the east component (west winds taken as negative in the computation) has a daily variation which corresponds fully to the daily variation of the valley winds: From 1100 to 2200 hours the east component is above normal, and at night and in the morning it is below normal. The daily variation of the Munich-Innsbruck pressure difference on these days corresponds to the above variation.

Thus it is possible without further information to sift out the influence of local winds with daily periods even in the presence of strong gradient winds, not only on warm summer days, but also during winter. The pressure gradient corresponding to the valley winds can always be measured, even quantitatively: For example, a Munich-Innsbruck pressure difference of 0.3, 1.0, 1.5 and 1.8 mm around 1400 hours corresponds to a maximum hourly value of the up-valley wind of 7, 12, 17 and 22 km/hr, respectively. If the daily pressure variation at Munich is approximately equated to that at the opening of the Inn valley into the Bavarian high plain (slightly north of Kufstein), the above-mentioned pressure differences are distributed over a stretch of around 80 km. The greatest pressure difference mentioned above would then correspond to a gradient of 2.4 mm per degree of latitude. The associated wind speeds are relatively low because of the great friction.

3. <u>The valley wind system as a circulation</u>

In contrast to the interpretation of J. v. Hann, the new theory regards the valley wind system not as a periodic flow alternating back and forward, but as a closed circulation. Only the results of E. Kleinschmidt [14] were available at first to support this view: Both at Lake Constance and at Friaul in the southern Alps, the midday balloons exhibited flow towards the Alps in the lower layers and descent from the Alps back to the plain at higher altitudes. This contrasts with the morning balloons. Admittedly, Kleinschmidt did not directly relate this observation to the actual valley winds, but postulated the existence of a larger circulation encompassing the Alps as a whole. During the day, this circulation would rise at the warmed outer slopes of the range and, above a certain neutral surface, would return to the plain.

Although Kleinschmidt's evidence may well have been too circumstantial, the study of A. Burger and E. Ekhart [34] left no doubt whatsoever about the existence of this closed circulation between Alps and plain: From numerous pilot balloons of a fairly large number of stations in the Alps, at their outer periphery and in a broader area around them, it was demonstrated with surprising consistency that the midday balloons, in contrast to the morning balloons, revealed that the air in the lower layers flowed towards the heart of the Alps, but above a neutral surface flowed back from the Alps on all sides towards the plain.

In more detail, however, this study also shows that the valley winds of the individual Alpine valleys also fit into this larger circulation system: During the day the air not only flows along the heated slopes at the periphery of the Alps, but in particular also penetrates through each valley into the interior of the Alps, where it rises as a lateral slope wind and as a "slope wind along the sloping valley floor" to above crest level, and then above a neutral surface turns back to the plain. Thus we can interpret the valley winds as local reinforcements of the large-scale thermal circulation between mountains and plain.

This upper-level return flow appears impressively from analysis of a large number of pilot balloons at Innsbruck [31] and at Trento [35]: Below the crest height, the air flows into both valleys during the day. At high altitudes, however, it flows northwards at Innsbruck but southwards at Trento, which lies south of the main Alpine ridge. The phenomenon occurs every day on average. In the case of Trento, the strength of this return flow, which for the most part must occur in the region of the Etsch valley itself (see p. 26), is quite considerable, i.e. 1-2 m/sec. As a mean value for the entire Alpine region, Burger and Ekhart [34] obtained only around 0.15 m/sec.

The studies cited seem fully to confirm that valley winds--in agreement with the new theory--can be interpreted as a closed circulation between valley and adjacent plain. Only at a few locations did checks of wind conditions indicate that v. Hann's notions (equalizing flow back and forward between two air masses at different altitudes) may be at least part of the explanation: A. Jelinek [22] came to this conclusion for the Mezza Selva station in the Assa valley, where he assumed an interplay between the flat uplands of the Sieben Gemeinden ("Seven Communities") at 1000 m and the Italian lowland plain. Burger and Ekhart [34, p. 350] reached an analogous conclusion for the daily wind over Vienna and Linz at 1 km altitude and, by way of explanation, assumed an equalizing flow between the Pannonian lowland plain and the Bavarian high plain.

4. Vertical extent of valley winds

According to the theory, the upper limit of valley winds whose development is unperturbed lies at that height at which unhindered pressure equalization in the horizontal direction is possible with the air above the plain. This applies in particular to longitudinal valleys, such as the Inn valley, in which the "effective crest altitude", i.e. the lower ridge if both lateral ranges have no perturbations in their ridgelines, represents the sharp and pronounced lower limit for such pressure equalization.

In cross valleys, on the other hand, e.g. in the Etsch or Sarca valleys, unhindered pressure equalization does not set in so directly at a well-defined altitude: The air flowing into the valley cannot escape over the lateral mountain crests of that valley, because valley winds are also developing in the adjacent parallel valleys. It is more likely that the upper-level air in the region of the valley itself has to flow back to the plain, unless the upper gradient wind participates in carrying away the rising air. The upper compensating flow here--as at Trento--must reach much greater strength than above longitudinal valleys. Moreover, greater uncertainty in the upper limit of the valley wind must be expected in such cross valleys. For example--as has already been mentioned, both the up- and down-valley winds above Trento reach much higher than would be expected from the elevation of the immediately adjacent ranges (according to L. W. Pollak [15], up to around 2000 m above sea level, whereas the immediately adjacent range can be estimated as approximately 1500 m high). In particular, the ridge of approximately 1500 m height separating the Etsch and Sarca valleys is not a factor in the upper limit of the valley wind. Rather, a valley-wind system encompassing both valleys seems to develop and to extend from the Brenta group in the west to the southern spurs of the Dolomites in the east.

Admittedly, it is also absolutely possible that the particularly high vertical extent is caused by the "larger valley wind" in Tollner's sense [13], i.e. the flow which Kleinschmidt claims blows in the lower layers from the plain into the heart of the mountain system.

An important observation is that the mere fact of favorable meteorological conditions for development of local winds (strong daily temperature fluctuation) is generally sufficient to prevent even strong upper-level winds from extending much below the crest altitude. If a valley is cut deep enough (e.g. Inn, Rhone or Etsch valley), the direction and strength of the upper gradient wind in itself only rarely influences the development of valley winds at the valley floor. This is understandable considering that the thermally induced pressure gradients responsible for periodic valley winds are very considerable at the floor of large valleys. For example, R. Billwiller, Jr. [36] found a pressure gradient reaching 4.0 mb per degree of latitude in the afternoon between two stations of the lower Rhone valley (40 km apart). This is far greater than the strongest gradients caused by general weather conditions in the Alpine region, except for purely local perturbations such as thunderstorms.

Since the valley wind gradients decrease with altitude and reach their zero value at the equalizing surface, while the general pressure gradients usually increase with altitude, there is a height at which both gradients have the same value. Obviously we can assume that this is the practical upper limit of the layer with valley wind. However, this altitude is not identical to that derived from the wind stratification. Only when the gradient wind has a direction opposite to that of the valley wind does the wind strength decrease to zero (where the two gradients maintain equilibrium). At higher altitudes, the strength of the wind increases again, but its direction is opposite.

In contrast, if the gradient wind has the same direction as the valley wind, a minimum wind velocity usually appears in the wind sounding. This minimum is utilized as a criterion for the vertical extent of the valley wind. As follows from Figures 1a and 1b, the minimum does not lie where the two pressure gradients are equal to each other, but where the thermal local pressure gradient decreases with altitude to the same extent as the general pressure gradient increases with altitude. In both Figures the light continuous curves represent the strength of the local valley wind considered separately; the broken curves represent the wind strengths caused by general weather conditions (corrected for friction and air density); and the bold continuous curves represent the sum of both wind strengths. The fact that this minimum is usually found below crest altitude is in

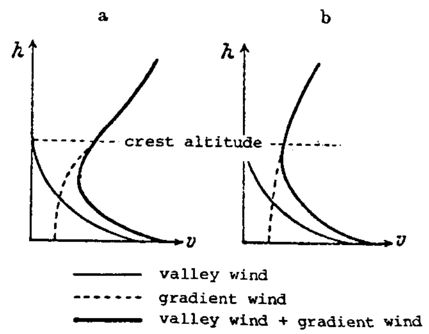

Figure 1: Wind minimum when valley wind and gradient wind have the same direction.

each case because the general gradient wind is already intensifying more strongly with altitude below crest altitude than the very weakened valley wind can still decrease with altitude. Thus the wind minimum indicates nothing more than the fact that the gradient wind considered separately becomes stronger than the local valley wind considered separately only above this altitude.

In the Ötz valley, Ekhart [17] believed that he could detect the beginning of a strong upper compensating flow even below crest altitude. It is cautioned, however, that a fairly strong gradient wind had approximately the same direction as this upper counter flow in all these cases. Thus an alternative explanation for the pilot curves could be that the gradient wind on these days reached far below crest altitude, but its direction was turned slightly in the valley direction. Thus it ought to be expected even in the Ötz valley that, during upper-level calms, the valley wind would develop up to the altitude of the lateral mountain crest.

At least it can be difficult to give a physical reason why thermally induced counter gradients should develop even below crest altitude. At best, it might be that the layer with above normal daily temperature fluctuation could not extend up to the lateral crest altitude due to glaciation at high altitudes. We would then have to assume that the glacier winds, which the investigations of Ekhart [17,37] and Tollner [13,38,39] show are most strongly developed in the afternoon, and which therefore cool the adjacent air by turbulent mixing during the afternoon, cause the temperature fluctuation in the uppermost layer of valley air to decrease below the normal value of the air above the plain (or above the main valley).

If the gradient wind blows across the valley direction, the valley wind virtually never turns gradually to the direction of the gradient wind, but rather does so with a sharp bend. Such a bend means that the wind shift is caused by a thermal interface: a gradient wind blowing across the valley direction can only extend below the elevation of the lateral crests if the air sinks behind one crest and rises again on the other side. As it descends, the air must become adiabatically heated. In general, therefore, when the temperature decrease with altitude is less than 1°C per 100 m, this air is warmer than the original valley air. In the presence of cross winds, therefore, the valley wind will be separated--not only during the night but also during the day--from the upper gradient wind blowing across the valley by a thermal interface, which lowers the internal friction and produces a sharp bend in the pilot balloon trajectories.

In a series of studies in which pilot balloons at Innsbruck were simultaneously released both at the valley floor and at various altitudes on the slope north of Innsbruck, the upper limit of the valley wind was found quite regularly to be depressed above the valley center, both for the nighttime down-valley wind and for the up-valley wind [40]. Ekhart [17] had previously made a quite analogous observation (only by day) by means of tethered balloons for the glacier wind of the Hintereisferner in the Ötz valley. Ekhart's suggestion at that time--that the lateral up-slope wind would suck the glacier wind flow somewhat upwards--is not consistent with the general phenomenon, because this occurs far above the range of the slope wind. Moreover, it exists not only during the day when the up-slope wind is blowing, but also during the nighttime down-slope wind. The coriolis force must be ruled out as an explanation even though it would compress the up-valley wind at an orographically left slope in the northern hemisphere, because the down-valley wind here would have to be much lower than above the valley center. An explanation with a thermal basis--as far as I can see--is equally unlikely: The daily temperature fluctuation in the region of the slope is probably greater than in the valley center, but such temperature differences would only cause cross circulation. They should not lead to reinforcement of the valley winds near the slope, since it is difficult to see how the daily temperature fluctuation on a horizontal along the valley center could increase more slowly up the valley than in the proximity of the slopes.

The most obvious explanation is that, in the presence of gradient winds blowing across the valley, their lower limit--and so the practically observable upper limit of the valley winds--is depressed above the valley center. In the cases observed by Ekhart [17] and in some cases studied by Jelinek and Riedel [40], this explanation may suffice. Characteristically, however, the valley wind above the slopes has been found to be more strongly developed than above the valley center if approximate calm (v less than 2 m/sec) prevails at higher altitudes. Thus we must probably assume that even such a weak gradient flow is depressed much more deeply above the valley center than in the vicinity of the lateral crests.

5. Gradients and thickness of the upper compensating flows

In order that a return flow of air--in daytime from the valley to the plain, and in nighttime from the plain to the valley--can develop above the effective crest altitude, the appropriate pressure gradients must exist. The problem of the origin of such thermodynamically induced gradients as well as the thickness of the upper return flow of a thermal circulation have been treated in general form elsewhere [41]. At this place it must suffice to relate only the essentials of those considerations to the valley winds.

The valleys with their lateral slopes represent a cold perturbation zone during the night and a warm perturbation zone during the day. This zone is surrounded by the unperturbed atmosphere above the plain. The upper limit of this perturbation zone is not at the altitude of the lateral crest, but extends somewhat beyond the crest due to turbulent exchange and radiative transfer. Thus pressure gradients which keep the upper compensating flow in motion develop above the pressure equalization surface (horizontal isobaric surface), which approximately coincides with the lateral crest altitude. The inclination of these surfaces of equal pressure increases with altitude above the pressure equalization surface, until the horizontal temperature gradients have vanished. If the temperature differences between valley and plain were equal to zero at all elevations above this thermal equalization surface, the upper return flow would have to reach to the outer limits of the atmosphere.

However, if it is required that the thermal circulation be confined <u>within</u> the atmosphere, then the horizontal pressure gradients which increase with altitude up to the thermal equalization surface must decrease again at still higher altitudes. This is possible only if there are temperature differences between valley and plain and if those differences are opposite to the conditions in the lower layers. <u>The necessary condition for the valley wind circulation to have an upper limit is therefore the existence of an upper warm</u>

zone above the lower cool zone of the valley during the night, and an upper cold zone above the lower warm zone during the day.

In the cited report, there is more detailed explanation of how such a peculiar temperature stratification (cold above warm and vice versa), which is also invariably found in other circulations, can come about. These discussions can be briefly repeated for the example of the up-valley wind: During the day, the air above the lateral mountain crest is warmed from below by turbulent mixing and possibly also by radiative transfer. This heat input decreases with increasing distance from the heat source, i.e. with elevation, and vanishes at some height H_1, which is not known closely.

The air above the valley--or more generally above the entire range--flows back at the upper levels on all sides towards the plain during the day. In the entire layer encompassed by this divergent flow, there must be a rising movement above the heat source. The rising movement must become weaker with altitude, partly because at the higher altitudes only that air rises which has not already been transported away horizontally in the lower layers, and partly because the horizontal cross section of the rising air flow becomes greater with increasing distance from the warm zone of the valleys. However, this rising movement cannot be regarded as a uniform movement in more or less parallel flow paths. Rather, the rising air mixes with the air flowing essentially horizontally as the gradient wind. The cooling caused by air rising in a stable stratified atmosphere is therefore greatest in the region of the lateral mountain crests. Because of mixing with other air masses as this air rises, the cooling effect diminishes with altitude until ultimately it practically vanishes at a certain height H_2.

Each air parcel above the lateral mountain crest is therefore subjected to heating from below on the one hand and to cooling due to rising movement on the other. The actual temperature stratification represents the resultant of the two influences. In particular, the warm zone of the valleys ends at that height H' at which the heating and cooling influences just maintain equilibrium. The height H_2 must basically extend further up than H_1, otherwise above H_1 the horizontal pressure gradients would remain unchanged with divergence, rising motions and cooling. In the layer between H_1 and H_2, therefore, only cooling is active, produced by the cold zone carried above the lower warm zone. The cold zone must also reach as high as the circulation on the whole, i.e. to that height at which the vertical motion also vanishes.

Observations demonstrate that there is an upper limit to the circulation of valley wind: Ekhart [31] concluded that the compensating flow above Innsbruck extended to 2-4 km on average, with maximum development of around 1/4 m/sec at around 3 1/2 km altitude. The pressure gradients which keep this upper return flow in movement must therefore intensify with altitude from the equalization surface (around 2 km) up to 3 1/2 km. At higher altitudes they must decrease again, reaching zero at around 4 km above sea level. However, this upper limit is certainly higher than 4 km on average for the Alps as a whole, since A. Burger and E. Ekhart [34] have found from combined observations of a fairly large number of stations that the Alps influence the daily winds up to at least 5 km.

The temperature distribution must therefore be such that the air above the valley exhibits a greater daily temperature fluctuation than that above the plain, even up to 3 1/2 km. Between 3 1/2 and 4 km (or 5 km), on the other hand, the conditions must be reversed, i.e. the air in this layer must be relatively cold above the warm zone of the valley during the day and relatively warm above the cold zone of the valley during the night.

As follows from the above discussion, the physical forces which produce such a contrasting temperature distribution are always present. If the air at individual locations rises particularly intensively above the valley (as up-slope wind), it is conceivable that such an air stream rises above its thermally induced equilibrium level and so in the uppermost part becomes colder than the surrounding air.

6. Daily variation of the valley winds

Whereas the wind strength at the plain reaches its daily maximum very soon after the sun has passed its zenith, at around 1300 hours, this maximum occurs considerably later for the up-valley wind. According to A. Defant [42], it occurs between 1500 and 1600 hours at Innsbruck on good summer days. According to Ekhart's climatic statistics for Innsbruck [43], it occurs around 1400 hours on average on all days during the summer months (much earlier in winter). The wind figures derived by Jelinek [22] for numerous stations in the Southern Tyrol also indicate that the maximum development of the up-valley wind occurs relatively late in large broad valleys. The average of beginning and end of the up-valley wind occurs around 1520 hours in the Etsch and Sarca valleys and around 1440 hours in the Suganer valley.

This late occurrence probably cannot be due to differences in the daily variation of turbulent exchange in the valleys as compared to the plain, but must be attributed mainly to the daily variation of thermal pressure gradients. In fact, satisfactory agreement is found between the daily variation of the valley wind and the daily variation of the pressure gradient between plain and valley. According to A. Defant [29], for example, the maximum of the pressure gradient between the Po plain and Bozen occurs around 1600 hours in spring and summer and around 1500 hours in winter and fall. F. Bondy's numerical data [30] for the daily variation of the Munich-Innsbruck pressure difference are also in agreement with this. For the May-August period, the pressure difference on average can be expressed by the equation:

$$\Delta b = 0.82 \text{ mm} \sin (220° + x) + 0.11 \text{ mm} \sin (17° + 2x).$$

The maximum gradient for up-valley winds occurs around 1515 hours if we consider only the first term.

According to R. Billwiller [36], the maximum pressure gradient for the up-valley wind occurs at 1500 hours in the lower Wallis. Another example (Death Valley) will be discussed later in another connection (p. 31). In all these cases, the maximum of the up-valley wind at the valley floor occurs much earlier than the afternoon minimum of air pressure. Thus the daily variation of air exchange with its maximum soon after the sun has passed its zenith can only partly compensate for the delay which normally occurs between the periods of force and the periods of motion.

Conditions in small valleys are different: the smaller the air mass to be set in movement, the more closely the development phase of the valley wind approaches that of the sun's altitude. In side valleys, the maxima of the valley wind sometimes occur several hours earlier than in a main valley. This situation produces interesting daily variations where side valleys open into a main valley. Such variations can be easily explained by superposition of the two out-of-phase wind systems (see [18]).

Two reasons can be given from the outset for the strong delay in larger as compared to smaller valleys. Firstly, as Ekhart [17, p. 251] expects, the delay in movement relative to the phase of the moving force is due to mechanical inertia, and so is greater the larger the air mass set in motion. Under such circumstances, however, the thermally induced pressure gradients presumably would reach their extreme values later the thicker the air layers whose different heating or cooling contributes to development of the pressure gradients. However, the thermodynamic influence of slope of the valley floor ought to be a decisive factor. This influence is explained on p. 36.

A closely related question is how the development phase of the valley wind changes with altitude above the valley floor. According to Ekhart [2], numerous balloons on good summer days at Innsbruck indicated "no or relatively slight shift in time of the beginning of valley wind with altitude", although the tendency would be to expect a considerable delay with altitude, since the temperature differences between plain and valley are initiated from below. In fact, if the pressure above the plain were to remain unchanged, we could assume with E. v. Everdingen [11] that there would be a delay of the daily variation of pressure gradients with height, because "the lowest layers are warmed earliest, even to intermediate altitudes in a very short time, whereas the highest layers are warmed only after circulation and exchange have worked their way upward". In reality, however, the conditions are more complicated, since the pressure gradients result from the difference between heating in the valley on the one hand and above the plain on the other, and since heating of the air above the plain is always delayed with altitude more strongly than in the valley.

Since the change in development phase of the valley wind with altitude can be directly derived only from individual daily series, it seems more expedient to evaluate the daily variation of pressure gradients between plain and valley from the admittedly scanty pressure recordings of high-altitude slope stations and valley stations: For the Hafelekar station (2265 m), the daily variation of air pressure, calculated from the months of May to August in 1929-31, is given in [10]. Even though Hafelekar is a crest station, it lies below the effective crest altitude, because both the central Alpine ridge in the south and the posterior chain of the Karwendel range in the north are higher, and prevent horizontal pressure equalization with the plain at the altitude of Hafelekar. The same article mentions the change in the variation of air pressure with altitude in the outer region of the Alps (Munich base station). The daily variation of the pressure difference between free atmosphere and Hafelekar is given by:

$$\Delta b = 0.08 \text{ mm} \sin (253° + x) + 0.03 \text{ mm} \sin (345° + 2x).$$

The phase angle of the first term in this equation exceeds the phase angle at Innsbruck by 33°, i.e. is more than 2 hours in advance. Admittedly, not too much importance should be attached to this numerical value, since it is not very accurate because of the low amplitude (less than

0.1 mm). In general, no stronger delay with altitude can be expected. Of special interest in passing is that--in agreement with the theory--the pressure gradients which keep the valley winds in movement can still be detected at the altitude of this crest station.

T. Zuchristian's analysis [44] of the daily variation of air displacement at Hafelekar on the basis of three years of recordings was also able to confirm the existence of the valley wind here: From 2100 to 700 hours, influence of the down-valley wind; rest of the time, influence of the up-valley wind. The existence of the valley wind at the altitude of Hafelekar can be deduced from the daytime leftwards veering of the wind vector, since Hafelekar lies at the crest of an orographically left slope where, according to A. Jelinek [22], leftwards veering can be expected due to the phase difference between slope wind and valley wind.

Another possibility of explaining the daily variation of pressure gradients at higher altitudes can be found from the Hochserfaus station (1940 m) in the upper Inn valley and the Vent station (1880 m) in the Ötz valley. At the first station, which is operated as a radiation research station by F. Haendel and is maintained by the Emergency Association for German Science, air-pressure recordings for a number of years are available. In contrast, the Vent meteorological station was only erected in the fall of 1935 by the German and Austrian Alpine Club. To obtain strictly comparable values, I have confined myself to the summer months of May to August 1936.

The first two rows of Table 1 give the terms of the sine series of the daily pressure variation for Vent and Hochserfaus. For direct comparison, Hochserfaus must be reduced to the level of Vent, which is 140 m higher. For the mean air pressure of Hochserfaus (612 mm) and the mean temperature of 8.5°C, the change in air pressure at the higher level corresponding to a 1°C change in mean temperature of the 140-m air column is 0.037 mm (in the same direction as the temperature change). Thus the daily variation of air temperature (row 3) yields the corresponding reduction values for the air pressure (row 4) and the air-pressure variation of Hochserfaus reduced to the level of Vent (row 5). The daily variation of pressure difference at the level of Vent can be represented by:

$$\Delta b = 0.29 \text{ mm} \sin (245° + x) + 0.01 \text{ mm} \sin (113° + 2x).$$

A phase angle of 245° for the first term corresponds to maximum development of the up-valley wind around 1345 hours.

No correction was made for the decrease in daily temperature fluctuation higher than Hochserfaus, since such a correction would be hardly perceptible. However, it could well appear that Hochserfaus does not lie at the outlet of the Ötz valley into the Inn valley, but rather 30 km upstream as the crow flies. In any case, the pressure conditions at the outlet of the Ötz valley (further down the Inn valley) more closely approximate those in the free atmosphere than do the conditions of Hochserfaus. Thus extreme values can develop such that the pressure variation of Vent is directly comparable with that in the free atmosphere (last row of Table 1). The maximum of the pressure gradient now occurs around 1400 hours. Thus the variation of pressure gradient which is critical for the valley wind in the Ötz valley at around 1900 m altitude lies between the values of the second-last and last rows. Admittedly this daily

Table 1

	p_1	q_1	p_2	q_2	a_1	A_1^0	a_2	A_2^0
1. Vent, b 0.01 mm	33.3	7.8	14.3	-13.6	34.1	76.8	19.9	133.6
2. Hochserfaus, b 0.01 mm	16.4	1.8	13.2	-15.1	16.5	83.7	20.0	138.8
3. Hochserfaus, t°C	-2.56	-1.62	0.61	0.28	3.03	237.7	0.67	65.4
4. Reduction, 0.01 mm	-9.5	-6.0	2.3	1.0	--	--	--	--
5. Hochserfaus, b reduced	6.9	-4.2	15.5	-14.1	8.0	121.3	18.2	132.3
6. Δb between Hochserfaus and Vent	-26.4	-12.0	1.2	-0.5	28.9	245.5	1.3	112.6
7. Δb between free atmosphere and Vent	-36.7	-20.8	1.9	2.0	42.1	240.5	2.8	43.5

variation of pressure gradient is not given for the floor of the Ötz valley, although it can probably be deduced from the existing pilot balloon studies that the strongest development of the valley wind at the valley floor occurs between 1300 and 1400 hours. Even though the air movement at the altitude of 1900 m will lag somewhat behind the daily variation of the pressure gradient, the phase delay of the valley wind with altitude appears to be only slight even here.

At this place we will again take up the question of why the air in the middle of the entire valley cross section exhibits a greater daily temperature fluctuation than the air above the plain. Because of the action of the lateral slope wind, thermal convection within the valley air is always greater than above the plain. Thus the heating effect from the slopes and valley floor is largely transferred also to higher layers at least up to crest altitude. Nighttime cooling acts similarly, not only at the bottom but also on the lateral slopes up to crest altitude. The consequence of all this is that the daily temperature fluctuation of the valley floor decreases more slowly with altitude than above the plain.

However, since the amount of sunlight falling on a given cross section is not greater above the valley than above the plain, it remains unclear why the daily temperature fluctuation in the middle of the entire air layer is greater than above the plain. Dynamic exchange is always primarily responsible: The strength of the gradient wind, even at the altitude of the lateral mountain crest, is considerably less than above the plain at the same altitude [31]. Consequently, the exchange which causes heat transfer between valley air and the higher layers of the free atmosphere is less intensive in the region of the valleys than above the plain. The heating or cooling occurring in the valley for local reasons remains largely confined to the valley air itself. Moreover--as already explained in the theory [1]--the mass of valley air in valleys with sloping sides is smaller than in valleys with vertical sides.

7. Winds of the Maloja wind type

According to the theory [1, p. 335], "winds of the Maloja wind type are possible at every pass altitude if the lateral mountain crests do not culminate at the pass altitude itself, but rise beyond this line in the direction of the descending valley". Only one influence--but in my opinion the most important--is considered in this context. Of course, it must be added that local features such as differences between the daily temperature fluctuations at the valley floor of both valleys, and differences in exposure, ground cover and open water can be significant factors in individual cases.

The reaction which the developing air flows exert on the original pressure field must also be taken into account. This influence, which is discussed in [1, p. 336], consists in the fact that not only are the orographic conditions at this location itself significant for the daily amplitude of the temperature of an entire valley cross section, but so also are "the conditions (daily fluctuation at the valley floor and relative altitude of the effective crest altitude) along the entire path that the air...has travelled." C. Braak [45] commented specially on this circumstance when he explained with respect to the true Maloja wind that the air blowing horizontally over the pass from the southwest had not had time to heat up correspondingly in the short path above the Bergell valley. Mainly for this reason, Braak believed that the valley cross section with the highest mean temperature and thus with the lowest afternoon air pressure at the valley floor lay east of the pass altitude.

It is not surprising in itself for the location of greater heating due to local features to lie not in the valley cross section of the pass altitude itself, but to be shifted towards one of the two valleys. However, this only explains that the air flows horizontally from the pass altitude to this location of lowest air pressure. In agreement with Ekhart [19], such winds can be called overlapping valley winds. For the true Maloja wind, however, it is also necessary that the air flow downwards in a certain layer thickness along the valley floor. In this process it must remain cooler than the displaced air despite the heating associated with compression. It is easy to see that a fairly steep gradient of the valley floor must put an end to such a downwards flow along the valley floor at once, if the flow of the "Maloja wind" is not considerably cooler than that of the normal up-valley wind of the opposite valley. Thus it is certain that the very slight gradient from Maloja pass eastwards is one of the main reasons that the Maloja wind blows even at the valley floor. It is also understandable that the conditions during the night (very stable air stratification) are much less favorable for development of the "Maloja wind," which during the night ought to flow upwards along the valley floor.

In contrast, overlapping valley winds, which blow only at high altitude above the pass line over into the other valley, ought to be much more frequent than the true Maloja wind. For example, Ekhart [19] finds that the valley wind of Kloster valley west of Arlberg blows even at the pass altitude from the same direction and in unvarying strength, but that at St. Anton in Stanzer valley, east of the pass, it generally is found only at altitudes higher than the pass altitude. Here, however, the slope from the pass height down to the Stanzer valley is very steep.

A. Jelinek [22] also found such an overlapping valley wind in the short Loppio valley. Because of this wind, the more strongly developed and probably also cooler (due to the influence of Lake Garda) up-valley wind of the Sarca valley sometimes blows over into the Etsch valley, albeit only at high altitude. At the valley floor, observations of ground-level winds at the outlet of the Loppio valley into the Etsch valley indicated a normal up-valley wind, opposite to the higher flow.

With respect to development of the overlapping valley wind, Ekhart also considered the possibility that it might result merely from inertial motion, in the sense that the more strongly developed valley wind would overlap above the pass altitude and flow against the gradient, until its kinetic energy had been consumed. This is certainly possible in individual cases, although this energy generally ought to be already consumed, at least near ground level. Pressure recordings from Maloja pass itself [46] show that the Maloja wind absolutely follows the pressure gradient.

Apparently the Maloja wind of the upper Engadin is not so simple in nature as was described in the older literature. As H. Klainguti-Schaumann [47] observed on the basis of three months of wind recordings at Celerina (19 km northeast of the pass altitude), the Maloja wind is absolutely not the typical good weather wind at Celerina, but rather the normal valley wind of the Inn valley (the "Bruescha") blows just as often here on good days. Comparison of the average weather conditions during the Maloja wind on the one hand and the Bruescha on the other make it probable that the general pressure gradient significantly influences development of the Maloja wind because of the exposed location of the upper Engadin.

Dr. Klainguti-Schaumann can also be thanked for her special interest in carrying out an aerologic investigation in summer 1937 with E. Moll, who prepared a preliminary report [48]. They found that a counter flow, which very probably can be regarded as the normal up-valley wind of the Inn valley, develops above the Maloja wind, which when it is fairly strongly developed along the valley floor flows downwards in a northeasterly direction to around Brail. The surface of separation between the air mass of the Maloja wind and that of the normal up-valley wind of the lower Engadin is not perpendicular. Rather, the Maloja wind--simultaneous temperature recordings show it to be a relatively cool flow--tapers off towards the northeast. At the surface of separation it gradually mixes with and becomes warmed by the warmer flow of the up-valley wind of the Inn valley, and ultimately rises together with this as the lateral slope wind. No special explanation is needed for the fact that the mean wind strength of the Maloja wind simultaneously also decreases from the pass altitude in a northeasterly direction, as the observations have shown. It appears that the Maloja wind is much more frequent at the pass itself than at Celerina, and so the upper gradient wind should not greatly affect the development of this wind at the pass itself, although it probably influences its range towards the northeast.

8. The valley winds in Death Valley

Judging from existing recordings, an interesting wind system with daily periods, which genetically is very similar to the Maloja wind, appears to exist in Death Valley, California. Death Valley represents a deep oval valley with a length of almost 300 km. The valley floor runs almost horizontally below sea level for a length of 120 km and a mean width of 10-12 km. The surrounding ranges become higher towards the north. The mean crest altitude is around 1500 m on the east side and around 2000 m on the west side. In the southern zone, the valley floor rises gently southwards up to the surrounding high plain, whose altitude is around 500 m. The northern zone of the valley rises steeply northwards. Figure 2 shows the contours at intervals of 2000 feet on the basis of the most recent 1:250,000 U.S. map. The 1000-foot contour is shown as a broken line where its position is of interest.

The brief climatic study of M. W. Harrington [49], based on the summer months of April to September 1891, also contains a diagram of the daily variation of air pressure and wind speed at the observation station (o). This variation in air pressure had been generally recognized by the fact that it exhibited the greatest daily fluctuation, where recordings were available at all:

$$\Delta b = 2.01 \text{ mm} \sin (352° + x) + 0.64 \text{ mm} \sin (151° + 2x).$$

This large daily fluctuation is evidence that the air temperature--not only near ground level but also in the entire valley cross section up to crest altitude of the lateral range--possesses an extremely large daily amplitude. According to the theory, however, the necessary condition for development of periodic valley winds is fulfilled by this fluctuation.

Figure 3 shows the daily variation of wind velocity (direction was not recorded) together with the daily variation of air pressure. A particularly striking feature is the late occurrence of the wind maximum (1700 hours) which, on the basis of existing studies in large Alpine valleys, is a characteristic sign of the valley wind

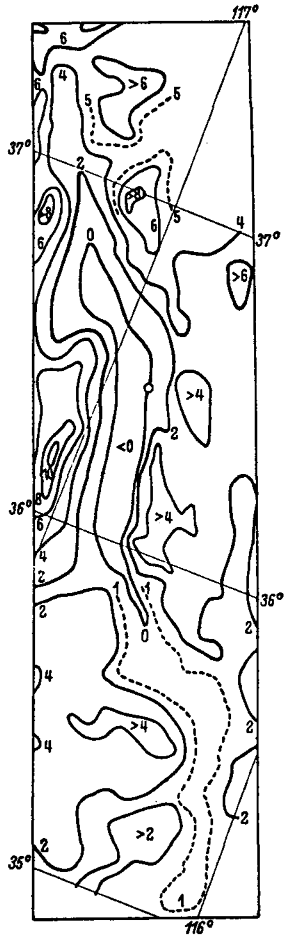

Figure 2: Death Valley. Contours at intervals of 2000 feet; o = position of the observation station.

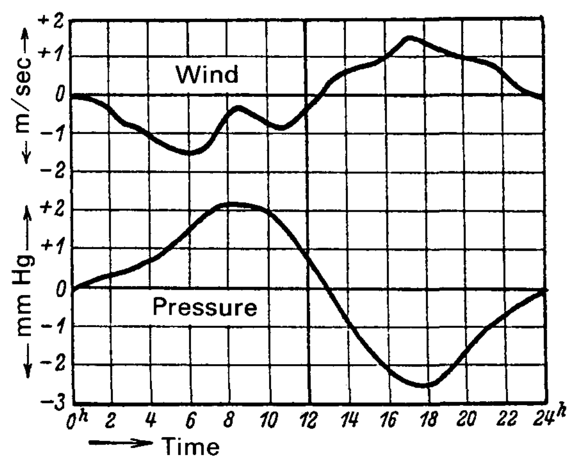

Figure 3: Death Valley. Daily variation of wind velocity and air pressure.

(since this maximum cannot be explained by a daily variation of air exchange, but only by a periodic fluctuation of the pressure gradient). The wind direction was determined only at two times of day (500 and 1700 hours). The wind was predominantly from the south, although the frequency of north winds was twice as high in the morning (27% north winds in the morning but only 14% in the afternoon).

Harrington himself suspected the existence of a wind system with daily periods [49, p. 28]: "There is also some evidence of the diurnal change of wind... This is a natural result of the heating and cooling of the higher valley to the north"). Examination of the daily curve of wind strength yields convincing evidence of the correctness of this supposition: The secondary maximum around 900 hours, which occurs in each of the five months of observation, probably can only be explained by a more or less regular change of direction around 1100 hours (secondary minimum). At this hour a valley wind entering from the south must start to develop. The second change of direction in the late evening (beginning of the wind blowing out of the valley) is always obscured by the frequent nighttime ground-level inversions. For this reason the mean wind strength continues to decrease during the entire night, and only increases around 600 hours, when the ground-level inversion is broken up by the effect of sunlight.

It is particularly noteworthy that the two wind maxima coincide with the two pressure extremes. Since the daily pressure variation in Death Valley is excessive, a local high-pressure zone must be present here around the time of the pressure maximum and a local low-pressure zone around the time of the pressure minimum, resulting in corresponding air movement.

This periodic wind system is particularly interesting from the theoretical standpoint, because the wind blowing into the valley must descend approximately 500 m from the surrounding high plain and so, despite the heating associated with compression, must arrive at lower temperature than the original valley air. Conditions during the night are analogous. The special orographic conditions indeed impair inflow of air during the day and outflow during the night, but by this fact the daily fluctuation of air temperature and so also the air pressure in the interior of the deep oval valley can continue to rise, until ultimately the intensified horizontal pressure gradient can provide the energy associated with vertical movement of air masses.

Since the "up-valley wind" in Death Valley must travel downwards for a long distance along the valley floor, and the "down-valley wind" must rise upwards, the conditions are very similar to those at Maloja pass. The slope of the valley floor to be overcome in Death Valley is also small, amounting to only 3 parts per thousand between the zero and 1000-ft contours. Even though--in my opinion--no further demonstration of the correctness of the new theory of valley winds is needed, it is worth pointing out that the existence of valley winds in a closed deep oval valley such as Death Valley cannot be explained by the theories of Hann and R. Wenger.

Since the valley winds in Death Valley are always caused by corresponding horizontal pressure gradients, it must be assumed that the amplitude of the air-pressure variation increases from south to north, despite the fact that the valley floor in the region of the station runs almost horizontally for a distance of

120 km. The reason for this increase lies in the rise of the lateral mountain ranges, i.e. of the effective crest altitude, from south to north. Even though the daily fluctuation of temperature at the valley floor probably should not increase from south to north, the daily fluctuation in the middle of an entire valley cross section certainly increases from the valley floor to the northwards crest altitude. This occurs despite the fact that air flowing into the south of the valley from the plain is warmed not only by convection, but also by compression on descending 500 m, provided the vertical temperature gradient is less than 1°C per 100 m.

9. Slope winds and their relationship to valley winds

The theory of valley winds attributes an important role to slope winds, since these largely provide the vertical air transport which is needed for development of the valley winds (since the actual valley wind--as [1] made seem probable--flows essentially horizontally, even in the presence of a sloping valley floor, at least when it is fully developed). Admittedly, no one has yet been able to obtain satisfactory numerical confirmation for this view. The only thing that can be asserted on the basis of double theodolite sightings to date is that no noteworthy difference in the apparent vertical speed of balloons can be observed in up-valley wind and down-valley wind, even in fairly steep valleys. Only near ground level must the valley winds naturally flow parallel to the valley floor, i.e. follow the gradient. Presumably, however, the gradient affects only the layer with the "shallow slope wind along the valley floor," which is usually distinguished from the actual valley wind even in the vertical distribution of wind strength. From the horizontal flow of the actual valley wind, it can be concluded that this does not represent a useful upwind for the glider pilot, a result which can be worth knowing when flying over the Alps.

The aerologic investigation of slope winds had to overcome some technical difficulties from the outset, since simple pilot balloon sightings were inadequate because of the occasionally considerable vertical flow components. Even double theodolite tracking failed when the layer being investigated was shallow. Thus, depending on circumstances, tethered balloons or balloons being sighted from two stations had to be used. The first preliminary study of E. Moll [18] itself revealed that the up-slope winds of the lateral slopes are only weakly developed in the lowest part, but that they increase in strength up to the crests. This observation was fully confirmed in the exhaustive investigations of A. Riedel [50], who further realized that the down-slope winds are shallow slightly below the lateral mountain crests but increase in thickness down towards the valley floor.

This observation necessitates a minor correction to the circulation diagram of the slope winds given in [1]: On average, air from the middle of the valley flows at all altitudes to the air mass of the up-slope wind, and only above crest altitude is there a flow on average back from the slope to the middle of the valley. Only when the uppermost part of the slope is in the shadow of cumulus clouds formed at low condensation level as a result of the up-slope wind does this part remain largely free of slope winds. These winds become separated from the slope, flow freely upwards as the "thermal upwind" and, under some conditions, can turn back towards the middle of the valley even below crest altitude. Similar conditions also exist with snow cover on the upper part of the slope. Nevertheless, traces of an up-slope wind can still be observed far above the snowline in such cases (see p. 39).

As for the up-slope wind, the flow of the down-slope wind must be supported in similar fashion at all altitudes by air from the center of the valley. The return flow to the valley center therefore occurs only in the lowest, cooled layer lying on the valley floor. Here, part of the air follows the slope of the valley floor and flows down the valley (down-slope wind along the valley floor), while the other part forces the valley air above the valley center upwards, thus causing mixing of the valley air. The only difference between the new diagram (Figures 4a and 4b) and the original drawings in [1] is that air flows at all altitudes towards the slope wind and that the compensating flow, which returns to the valley center, occurs only above crest altitude in the case of the up-slope wind and only in the lowest layer of air lying on the valley floor in the case of the down-slope wind.

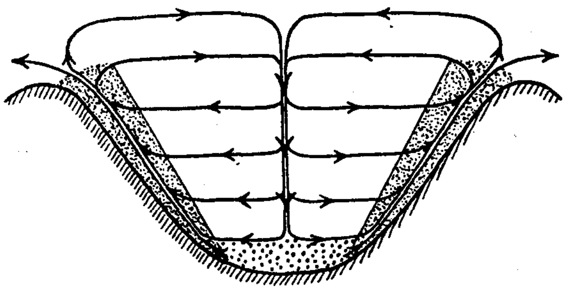

Fig. 4a.

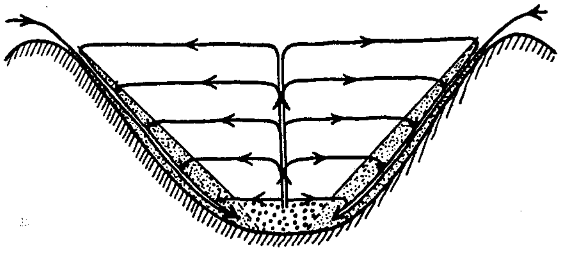

Fig. 4b.

Figure 4: Diagram of the topographic winds. a) In the presence of up-valley wind; b) In the presence of down-valley wind.

This modification to the diagram was also necessary for another reason: Analysis of the daily variation of air displacement in various valleys of South Tyrol by A. Jelinek [22] has led to the result that, in precisely those narrow V-shaped valleys in which winds exclusively in the valley direction would have been expected, broad daily wind roses were obtained. In other words, cross winds--always from the shaded slope to the sunlit slope--were observed whenever the valley winds passed through the zero value. Thus the slope winds exist right down to the valley floor in narrow valleys. The broad wind rose then results from the fact that, firstly, the slope winds react more rapidly to changing heating conditions because of the much smaller air mass participating in the flow and, secondly, the location of most intense radiation generally migrates from one slope to the opposite slope between morning and afternoon because of the change in the sun's altitude.

From the new diagram it is now understandable that, in valleys with broad valley floor, the slope winds in the region of the valley floor can exist only at the foot of the slopes, whereas its influence is imperceptible towards the middle of the valley. Thus the daily wind roses at stations in broad valleys are elongated in the valley direction, although it would have been thought that there was more room for development of cross winds in broad valleys.

This diagram is now consistent with the extremely graphic description which C. Braak gives of the development of slope winds in Central Java: "The movement of the cumulus clouds shows that the air movement is directed towards the mountains even at cloud level," and further: "the rising valley wind (correctly up-slope wind) is readily perceived even here (at the summit), although the cloud patterns usually separate from the mountainside before they reach the summit, rise more vertically upwards and at their highest parts bend back in the direction of the plain" (my underlining).

In the same article, Braak cited an example in which the nighttime ascent of air above the middle of the valley (see Figure 4b) became directly visible by condensation (Palu valley at Celebes): "The mountain slopes are usually shrouded in clouds during the day, whereas blue sky can be seen above the valley at Palu. Here it rains most frequently during the night or in the early morning...At night, the air above the valley is probably forced upwards in this long channel (Palu valley + Palu bay) by mountain winds descending along the slopes (i.e. down-slope winds), so that condensation is favored right above the valley."

Just as at the valley floor in narrow valleys, an interplay between valley wind and slope wind has always been found at exposed slope stations: Soon after the slope becomes sunlit, the up-slope wind develops, although it usually diminishes before sundown, when ground radiation outweighs the effects of insolation. In the intermediate period up to the beginning of the down-slope wind, the up-valley wind, whose development relative to the slope wind is more or less delayed depending on valley size, sweeps up to the slope. In quite analogous fashion, the down-valley wind blows in the morning between the disappearance of the down-slope wind and the beginning of the up-slope wind. On this basis, A. Jelinek [22] proposed the theorem that the wind vector rotates during the day describing a clockwise circulation at orographically right slopes and a counter-clockwise circulation at orographically left slopes.

The influence of the slope wind on the valley wind has been repeatedly observed in not too broad valleys by pilot balloons released from the middle of the valley, specifically at those times of day when the up-valley wind changes to the down-valley wind and vice versa. A cross circulation extending beyond the middle of the valley has frequently been observed under such conditions. Air from the middle of the valley flows towards the sunlit slope and returns to the valley center at higher altitudes in this circulation.

Strictly speaking, the diagram shown in Figures 4a and 4b, as well as the theory [1], are valid only for the time of full development of the two wind systems, i.e. in the early morning and after midday. However, since the development phases of slope winds and valley winds are shifted with respect to each other, the interaction between the two wind systems changes in the course of the day. Thus it is necessary to attribute this daily change as well to thermodynamic principles.

Although the action of the up-slope wind begins in the morning, after sunrise, with heating of the slopes, the valley air as a whole is cooler than the air above the plain. Thus the pressure in the valley is higher than that above the plain, and so the down-valley wind is still active. At the altitude of the lateral mountain crests, there is still no reason for the slope air which has risen to return to the plain. Rather, the beginning up-slope wind must turn back completely towards the middle of the valley. Moreover, the downwards movement which is necessary on average to maintain the down-valley wind must blow with such intensity above the middle of the valley that it more than compensates for the upwards movement along the slopes. This is the period during which scattered cloud cover above the valley breaks up, while horizontal cloud banks form at the slopes, particularly after rainy days. Because the valley air sinks as a whole, it is now warmed more strongly than the air above the plain, and the horizontal pressure gradients weaken until ultimately the valley air has the same temperature as the air above the plain.

When the down-valley wind then passes through the zero point and becomes the up-valley wind, the circulation of slope wind is then equalized, i.e. equally much air is conveyed up along the slopes as sinks again above the middle of the valley. This is the period of most rapid warming of the valley air as a whole. The more the up-valley wind then increases in strength, the more the vertical downwards movement above the middle of the valley diminishes compared with the up-slope wind, which must to an increasing extent supply the upper compensating flow of the up-valley wind until ultimately, when the up-slope winds in the afternoon have already become weaker due to decreased radiation, the air flowing upwards along the slopes is entirely returned to the plain above the crest altitude. The vertical mixing of valley air and so the descending movement above the middle of the valley have thus ceased, and the temperature of the valley air has reached its maximum, provided direct heating by solar radiation can be disregarded.

Approximately at the time of this temperature maximum, the pressure gradient directed into the valley also reaches its greatest value. In the later afternoon, when the up-slope winds have weakened still more, they are no longer capable alone of providing the upwards transport required by the strength of the up-valley wind. Thus the valley air must to a progressively increasing degree exhibit a vertical component in its entire mass, until ultimately, when the up-slope winds cease around sunset, the valley air must flow upwards in its entire layer parallel to the valley floor.

In the evening, when the down-slope winds have begun, the upwards movement above the middle of the valley must be so great that it overcompensates for the downwards movement along the slopes. The result--provided the air is stably stratified--is more rapid cooling than above the plain, so that the temperature difference between valley and plain and simultaneously the pressure difference between plain and valley reach the zero value.

While cooling of the valley air due to vertical upwards movement above the middle of the valley progresses, the down-valley wind begins. As it gains strength, the air descending from the slope is conveyed to a progressively increasing degree out of the valley, so that progressively less remains for the air stream rising above the center of the valley, until ultimately--at the time of the temperature minimum of the valley air--the vertical movement above the middle of the valley ceases. When the down-slope winds then cease in the morning, the valley air in its entire layer thickness parallel to the valley floor flows down the valley and thus heats up more rapidly than the air above the plain.

It is immediately obvious that break-up of the valley-wind gradients after the slope winds have ceased occurs more slowly the smaller the slope of the valley floor. Thus it can be deduced that the phase delay of the valley winds relative to the slope winds is greater the less the gradient of the valley floor. This is consistent with the observations that the small valleys investigated, with their smaller phase delay of the valley winds, also have more strongly sloping valley floors in all cases. The latest occurrence of the wind maxima has been found in Death Valley, i.e. in that case in which the valley floor of a completely closed deep oval valley rises towards both sides.

The aerologic investigation of slope winds by A. Riedel [50] and A. Jelinek [52], in addition to providing general confirmation for this theoretically expected interplay between slope wind and valley wind, also yielded a clearer explanation of the daily development of the slope wind in terms of strength and thickness and in terms of the influence of the immediate environment. All individual results obtained in these investigations can without difficulty be understood from the viewpoint that there is a circulation between the air near the slopes, which is warmed by insolation or cooled by ground radiation, and the less influenced air at greater distance from the slope. For example, it was found that the up-slope wind begins earlier at an eastern slope than at a southern slope and that, according to [27], the beginning of the up-slope wind migrates with the shade boundary. It is also immediately clear that every furrow in which the temperature difference with respect to the noninfluenced air is increased reinforces the slope wind. Moreover, every cloud which casts a shadow on part of the slope brakes the up-slope wind, and finally, the up-slope wind does not blow over the slope in one continuous stream, but every irregularity of the terrain produces turbulent mixing with the cooler outside air. It can also be understood why the up-slope wind does not reach its maximum strength soon after noon on days with slight cumulus formation at the mountain crests, but instead a secondary minimum of wind strength forms around this time and simultaneously a minimum vertical thickness of the up-slope wind can also be observed.

The layer thickness of the up-slope wind, measured normal to the smoothed slope surface, extends to around 300 m under favorable conditions in the upper parts of the slopes of the Inn valley. Along the vertical, the layer thickness would be correspondingly more, depending on the gradient of the slope. On the basis of cloud observations, C. Braak [51] suggests a thickness "which certainly exceeds 500 m" for Java. This fairly great thickness of the slope wind in Java ought to be due--apart from the greater daily temperature fluctuation in the tropics--mainly to the circumstance that Braak's observations dealt with air exchange between heated slopes and the free atmosphere, whereas in the Inn valley even the valley air has an above normal temperature amplitude.

According to R. Wenger's theory, the horizontal temperature gradient from the slope to the outside air is very important for development of the slope wind. For the same radiation conditions, therefore, the slope winds even in the region of the Alps must be developed more strongly the smaller the temperature fluctuation at some distance from the slope. In narrow valleys, therefore, weaker slope winds can be expected than in large, broad valleys. In contrast, the strongest development is expected at the outer periphery of the Alps where the air at some distance from the slope has only a very slight daily temperature fluctuation. This is particularly true for fairly high relative altitudes, at which the daily period of the temperature in the free atmosphere has almost vanished, whereas the slopes still exhibit a strong daily fluctuation. This observation may also be useful for glider flights in the mountains.

Jelinek's temperature measurements [53] at slopes near Innsbruck using cableway towers have provided satisfactory confirmation of Wenger's theory of slope winds: Strongly superadiabatic temperature gradients always prevail in the presence of the up-slope wind, and ground-level inversions in the presence of the down-slope wind. With the change of temperature gradient from increase to decrease with altitude and vice versa, the direction of the slope wind also changes. These measurements also confirm a suspicion which T. Zuchristian [44] had already expressed on the basis of three years of wind observations at three times of day at the Seegrube on the one hand and at the Patscherkofel on the other: Even on snow-covered slopes there is regularly an up-slope wind on sunny days. This is due to the lower-lying forest, from whose tops the snow immediately falls and which then act as a dark heating surface. This influence reaches up to the Hafelekar crest station.

The thermal stratification of slope winds of the smallest extent was already investigated earlier by H. Ungeheuer [54] in the Taunus and by H. Schultz [21] in the Wisper valley. Admittedly, these studies dealt only with microclimatic phenomena, which essentially appear to be influenced by the type of vegetation on the slopes.

10. Local perturbation of the valley winds at Trento

It still remains to be proved that the abnormal development of the valley winds at Trento does not contradict the correctness of the new theory of valley winds, but that a local perturbation is involved here. The existing statistics on frequency of wind direction at the three observation times in the Etsch valley [43, Table 14] already show that the development of the valley winds is completely normal south of Trento (at Ala) and also north of Trento (St. Michele), but that at Trento itself the up-valley wind rarely blows around 1400 hours. Defant [43] found up-valley wind in summer around 1400 hours on 19.9 days per month at Ala and 17.2 days at St. Michele, but only 3.5 days at Trento.

The mean pressure fluctuation between 700 and 1400 hours at various stations in the Etsch valley is even clearer proof of a local perturbation in the vicinity of Trento. The existence of valley winds requires that the air pressure along a horizontal decrease into the valley during the day and increase during the night, and so the daily pressure fluctuation reduced to constant level also increases into the valley. As the following numerical values compiled by Schaller [35] show (according to observations during May to August in 1898 to 1902), this pressure fluctuation indeed increases from Ala to Trento between 700 and 1400 hours, and then decreases as far as St. Michele, from where it again increases into the valley.

	700-1400 hr, mm
Ala	0.92
Rovereto	1.35
Trento	1.62
St. Michele	1.40
Gries	2.01

Thus countergradients can be expected between Trento and St. Michele, although it cannot be asserted on the basis of the above numerical data whether this perturbation influences the down-valley wind around 700 hours or the up-valley wind around 1400 hours or both. The wind statistics, however, prove that the down-valley wind is normally developed at Trento, and only the up-valley wind appears to be perturbed. In any case, the pressure fluctuations are consistent with the abnormal lack of up-valley winds at Trento.

According to Schaller [35], the up-valley wind at ground level blows only from 1100 to 1800 hours even on good days in summer (May to August), whereas at 1000 m it blows until around 2400 hours. During the transition months of March, April, September and October, there is no south wind at all on average at ground level, although this probably does blow at 100 to 200 m above the ground. Thus the perturbation primarily influences the layers near the ground. The obvious reason to suspect is the inflow of relatively cold air. In actual fact, the orographic conditions make such inflow very likely: Barely 2 km north of Trento there is a low pass (440 m relative altitude) between the Etsch and Sarca valleys. The Sarca up-valley wind, which blows over Lake Garda, always remains cooler than the Etsch up-valley wind and to some extent flows over the pass into the Etsch valley, which is still broadly developed northwards. Because of its lower temperature, this flow descends during the day, if the vertical temperature gradient is large, all the way to the floor of the Etsch valley, where it flows not only northwards as a cold air mass, but also some distance southwards, where it forms a cold-air wedge above which the normal Etsch up-valley wind must rise.

In the first phase of the up-valley wind, therefore, it blows only at upper levels at Trento. It develops at ground level only when it is sufficiently strong to force the wedge of cold air northwards. This explains the observation of L. W. Pollak [16] that "the north winds break up from above during the course of the day."

This cold-air wedge, which is continuously transferred into the warmer airflow of the Etsch up-valley wind with turbulent mixing at its front edge and upper boundary, is probably also responsible for the fact that the wartime airfield, originally located north of Trento, had to be moved because of landing difficulties.

In reviewing the results of existing investigations on the valley wind, we can readily assert that this wind system not only has been fundamentally explained, but has already been researched in so many details that valley winds now represent one of the best-known wind systems. This circumstance--apart from the fact that valley winds are relatively easy to study because of their spatial limitations--is primarily due to the aerologic research methods, which best permit a spatial approach to analysis. The Emergency Association for German Science, which enabled the numerous individual studies of the Innsbruck Institute by providing sufficient grants, is due most of the credit for this advance. The significance of the results summarized in the present article is not only in the explanation of a local phenomenon, but it can rather be hoped that, starting from this basis, other thermal circulations can be interpreted more clearly than has been possible until now.

REFERENCES

1. A. Wagner, Neue Theorie der Berg- und Talwinde [New theory of mountain and valley winds]. Meteor. Z. 49 (1932) 331.

2. E. Ekhart, Zur Aerologie des Berg- and Talwindes [Aerology of the mountain and valley wind.]. Beitr. Physik fr. Atm. 18 (1932) 1.

3. E. Ekhart, Weitere Beitraege zum Problem des Berg- und Talwindes [Further contributions to the problem of the mountain and valley wind]. Beitr. Physik fr. Atm. 18 (1932) 242.

4. E. Ekhart, Die Tageszeitenwinde der Alpen. Eine Darstellung nach dem neuesten Stande unseres Wissens. [The daily periodic winds of the Alps. A representation based on the latest status of our knowledge]. Naturwiss. (Im Druck.)

5. A. Wagner, Hangwind - Ausgleichsstroemung -- Berg- und Talwind [Slope wind -- equalizing flow -- mountain and valley wind]. Meteor. Z. 1932, 209.

6. I. Fournet, Des brises de jour et de nuit autour de montagnes, [Daytime and nighttime winds around mountains]. Uebers. in Poggendorfs Annalen. Erg.-Bd. I. 1842.

7. R. Wenger, Zur Theorie des Berg- und Talwindes. [Theory of the mountain and valley wind]. Meteor. Z. 1923, 193.

8. Vgl. J. v. Hann, Lehrbuch der Meteorologie. 3. Aufl. [Textbook of Meteorology, 3rd Edition]. 446.

9. M. Saigey. Influence des inégalités du sol sur l'equilibre de l'Atmosphère [Influence of ground irregularities on the equilibrium of the atmosphere]. Petite Physique du Globe. Paris 1842.

10. A. Wagner, Der taegliche Luftdruck- und Temperaturgang in der freien Atmosphaere und in Gebirgstaelern [The daytime air-pressure and temperature variation in the free atmosphere and in mountain valleys]. Gerl. Beitr. Geophys. 37 (1932) 315.

11. E. v. Everdingen, Zur Theorie der Berg- und Talwinde [Theory of mountain and valley winds]. Beitr. Physik fr. Atm. 19 (1932) 109.

12. L. Barda, Sommerliche Pilotbeobachtungen im Suedwesten Osterreichs waehrend der Kriegszeit [Summertime pilot observations in southwest Austria during the war]. Ref. v. A. Schedler. Meteor. Z. 1928, 131.

13. H. Tollner, Berg- und Talwinde in Österreich [Mountain and valley winds in Austria]. Beiheft z. Jb. 1928 d. Zentr. Anst. f. Met. u. Geod. Wien. 1931.

14. E. Kleinschmidt, Der taegliche Gang des Windes in der freien Atmosphaere und auf Berggipfeln [The daily variation of the wind in the free atmosphere and at mountain tops]. Beitr. Physik fr. Atm. 1922, S.1.

15. L. W. Pollak, Berg- und Talwinde im Becken von Trient [Mountain and valley winds in Trento basin]. Meteor. Z. 1924, 18.

16. V. Wiener, Die Windrichtungen im Becken von Trient [The wind directions in Trento basin]. Diss. Prag 1929. Ref. in "Lotos" 77 (1929) 16.

17. E. Ekhart, Neuere Untersuchungen zur Aerologie der Talwinde: Die periodischen Tageswinde in einem Quertal der Alpen. [New investigations on the aerology of valley winds: The periodic daily winds in an Alpine cross valley]. Beitr. Phys. fr. Atm. 21 (1934) 245.

18. E. Moll, Aerologische Untersuchung periodischer Gebirgswinde in V-foermigen Alpentaelern [Aerologic investigation of periodic mountain winds in V-shaped Alpine valleys]. Beitr. Physik fr. Atm. 22 (1935) 177.

19. E. Ekhart, Die Windverhaeltnisse des Arlbergs. II. Teil: Aerologische Beobachtungen [Wind conditions of the Arlberg, Part II: Aerologic observations]. Gerl. Beitr. Geophys. 49 (1936) 7.

20. H. Barschall, Über die Gebirgswinde in den mittleren Vogesen, [Mountain winds in the central Vosges]. Meteor Z. 1919, 137.

21. H. Schultz, Über Klimaeigentuemlichkeiten im unteren Rheingau mit besonderer Beruecksichtigung des Wisperwindes [Climatic peculiarities in the lower Rhine district, with special analysis of the Wisper wind]. Frankfurter geogr. Hefte 7 (1933). H. 1.

22. A Jelinek, Untersuchung periodischer Tageswinde in Suedtirol [Investigation of periodic daily winds in the southern Tyrol]. Beitr. Physik fr. Atm. 21 (1934) 223.

23. F. Lauscher und O. Eckel. Pilotierungen im Gailtal (Kaernten) [Pilot observations in the Gail valley (Carinthia)]. Meteor Z. 1934, 201.

24. W. Schwabl, Zur Kenntnis der Beeinflussung der allgemeinen Stroemung durch ein Gebirgstal [Knowledge on the influence of a mountain valley on the general flow]. Meteor. Z., 1934, 342.

25. E. Ekhart, Bemerkung dazu [Remark on (24)]. Meteor. Z, 1934, 477.

26. W. Schwabl, Erwiderung, [Reply to (25)]. Meteor. Z., 1934, 479.

27. A. Roschkott, F. Steinhauser und F. Lauscher, Winduntersuchungen im Sonnblickgebiet [Winter investigations in the Sonnblick region]. 42. Jb. d. Sonnblickvereins 1933, 15.

28. Vorlaeufiger Bericht in: H. Kinzl, Die Cordillera-Blanca-Expedition 1936 [The Cordillera Blança Expedition 1936]. Mitt. d. D. u. O. A.-V. 1937, Aprilheft.

29. A. Defant, Der Innsbrucker "Schoenwetterwind" [The Innsbruck "fair weather wind"]. Ber. d. naturw.-med. Vereins Innsbruck 30 (1905/06).

30. F. Bondy. Über Beziehungen zwischen periodischen Talwinden und verschiedenen meteorologischen Faktoren [Relationships between periodic valley winds and various meteorological factors]. Diss. Innsbruck 1935.

31. E. Ekhart, Die Windverhaeltnisse ueber einer inneralpinen Station [Wind conditions above an inner-Alpine station]. Beitr. Physik fr. Atm. 23 (1936) 165.

32. R. Kanitscheider, Beitraege zur Mechanik des Foehns [Contributions to the mechanics of the foehn]. Beitr. Physik fr. Atm., 18, 27.

33. R. Kanitscheider, Beitraege zur Mechanik des Foehns, II [Contributions to the mechanics of the foehn, II]. Beitr. Physik fr. Atm. 24 (1936) 26.

34. A. Burger und E. Ekhart, Über die taegliche Zirkulation der Atmosphaere im Bereiche der Alpen [The daily circulation of the atmosphere in the region of the Alps]. Gerl. Beitr. Geophys. 49 (1937) 341.

35. E. Schaller, Aerologische Untersuchung des periodischen Talwindsystems von Trient. [Aerologic investigation of the periodic valley-wind system of Trento]. Diss. Innsbruck 1936.

36. R. Billwiller, Der Walliser Talwind und der taegliche Barometergang [The Wallis valley wind and the daily barometric variation]. Meteor. Z. 1915, 423.

37. E. Ekhart, Über einige Gletscherwind-Messungen in den Ötztaler Alpen [Some measurements on glacier winds in the Otz valley Alps]. Zeitschr. fr. Getscherkunde 22 (1935) 217.

38. H. Tollner, Gletscherwinde in den Ostalpen [Glacier winds in the eastern Alps]. Meteor. Z. 1931, 414.

39. H. Tollner, Gletscherwinde auf der Pasterze [Glacier winds on the Pasterze]. 44. Jb. d. Sonnblickvereins 1935, 38.

40. A Jelinek und A. Riedel, Über die Schichtdicke der periodischen Lokalwinde im Inntal [Layer thickness of the periodic local winds in the Inn valley]. Beitr. Physik fr. Atm. (Im Druck.)

41. A. Wagner, Über die vertikale Maechtigkeit einer thermischen Zirkulation und die vertikale Temperaturschichtung in derselben [Vertical thickness of a thermal circulation and vertical temperature stratification in that circulation]. Gerl. Beitr. Geophys. (Im Druck.)

42. E. Ekhart, Klima von Innsbruck [Climate of Innsbruck]. Ber. d. naturw.-mediz. Vereins Innsbruck 43/44 (1931/32, 1933/34).

43. A. Defant, Berg- und Talwinde in Suedtirol [Mountain and valley winds in southern Tyrol]. S.-B. d. Akad. Wiss. Wien, 118, Abt. IIa (1909) 553.

44. T. Zuchristian, Ergebnisse der Windregistrierungen auf einer Kammstation (Hafelekar, 2265 m) [Results of wind recordings at a crest station (Hafelekar, 2265 m)]. Meteor. Z. (1935) 177.

45. C. Braak, Der Malojawind [The Maloja wind]. Meteor. Z. 1933, 231.

46. R. Billwiller, Der Talwind des Oberengadins [The valley wind of the Upper Engadin]. Meteor. Z. 1896, 129.

47. H. Klainguti-Schaumann, Über die Windverhaeltnisse des Engadins, speziell den Malojawind [Wind conditions of the Engadin, especially the Maloja wind]. Meteor. Z. 1937, 289.

48. E. Moll, Studien zur Frage des Malojawindes [Studies on the question of the Maloja wind]. Vortrag b. d. Tagung der Deutschen Meteorologischen Gesellschaft, Frankfurt a. M., Okt. 1937. Meteor. Z. 1937 (Im Druck).

49. M. W. Harrington, Notes on the Climate and Meteorology of Death Valley, Cal. U.S. Weath. Bur. Bull. Nr. 1, Washington 1892.

50. A. Riedel, Ergebnisse von Doppelvisierungen im Innsbrucker Nordkettengebiete [Results of double theodolite pilot balloon sightings in the north range of Innsbruck]. Diss. Innsbruck 1935.

51. C. Braak, Der Talwind [The valley wind]. Gerl. Beitr. Geophys. 32 (1931) 83.

52. A. Jelinek, Beitraege zur Mechanik der periodischen Hangwinde [Contributions to the mechanics of the periodic slope winds]. Beitr. Physik fr. Atm. 24 (1937) 60.

53. A Jelinek, Über den thermischen Aufbau der periodischen Hangwinde [Thermal development of the periodic slope winds]. Beitr. Physik fr. Atm. 24 (1937) 85.

54. H. Ungeheuer, Mikroklima in einem Buchen-Hochwald am Hang [Microclimate in a high beech forest on the slope]. Biokl. Beibl. 2 (1932) 75.

CONTRIBUTIONS TO ALPINE METEOROLOGY

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CONTRIBUTIONS TO ALPINE METEOROLOGY

Dedicated to the Memory of Arthur Wagner

E. Ekhart

Introduction

In September of 1941 an artillery school held maneuvers in the Alps. During the course of these maneuvers, numerous meteorological-aerological measurements were taken. These measurements were intended first and foremost for the specific requirements of the maneuvers, but they also represent basic material well suited in extent and volume for purely scientific studies.

In particular, several series of radiosonde measurements of upper air winds and temperatures are applicable to the study of local mountain wind systems. If this leads to a renewed, broader interest in and fresh insights into alpine diurnal winds, then I have the leaders of the maneuver to thank. I had the honor of functioning as a technical advisor on questions of alpine meteorology during part of the maneuvers. The leaders of the maneuvers, with the greatest understanding for the project, placed all the necessary supporting data at my disposal. Therefore, it is a pleasure to express again my thanks to the military leadership, above all to the commander, Captain Froben, for the considerable cooperation with which they responded to my wishes. Recognition is also due the leadership of the meteorological service, which was in the hands of Mr. Macht and then of Mr. Kaiser and Mr. Vetter. The collection of meteorological data in the mountains was often technically difficult. The success of the field program is due to the tested organizational talent of these men and to their careful planning. I owe special thanks to Mr. Macht, who worked closely with me during the first three series of maneuvers to prepare and carry out the meteorological work. He also helped with the general planning of the meteorological program for the later maneuvers. In so doing, he took into consideration the special requests resulting from the planned scientific use of the observations.

The maneuvers encompassed the most varied landscapes: longitudinal and cross valleys [Ed. Note: Valleys oriented parallel to or perpendicular to the longitudinal axis of the Alps, respectively], broad and narrow valleys, basins and mountain passes. This opened the most tantalizing and varied perspectives for the research of alpine flows.

The bulk of the observations consisted of series of frequent upper air wind measurements for two or more days at three locations within the maneuver area. Two stations were equipped with double theodolite balloon systems, the third with a single theodolite, which was used in part for optical tracking of radiosonde ascents. In addition, continuous quarter-hourly to half-hourly ground measurements of wind, temperature, and weather conditions were taken at the three stations, which were manned by specially equipped weather observers. In all cases, the abundance of measurements has made it possible to interpolate between observations, as was necessary to deduce the characteristic daily variations of the various parameters.

The analysis of all the observations produced a wealth of results. These results not only support and expand our theoretical and empirical understanding of the nature and genesis of the Alpine diurnal wind systems, but also provide new insights into the structure of the mountain atmosphere. Future, more detailed studies of the mountain atmosphere will be indebted to the 1941 field program.

The significant results of the study will be presented in parts, corresponding to the individual stages of the maneuvers. A brief description of the geography, supplemented by maps and diagrams, will make it easier for the reader to visualize the observational area as it relates to to the particularly interesting question of local wind conditions.

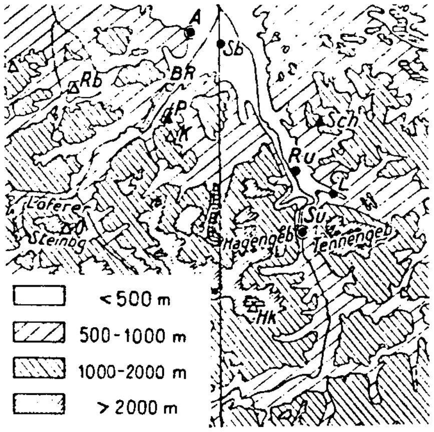

Figure 1. The research area with the aerologic measurement stations. Su = Sulzau, Ru = Russeck, L = Lehngries, Sb = Salzburg airport. [Ed. Note: Ekhart's Figure 1 could not be reproduced, so we have substituted a portion of Figure 1 from his 1948 paper.] "

AEROLOGY OF THE VALLEY WIND SYSTEM IN THE SALZACH VALLEY AND THE LAMMER VALLEY

Summary

An analysis of series of measurements of upper air winds in the Salzburg Alps has resulted in new knowledge of small scale air flow in mountain valleys. The main results are a longitudinal section scheme of valley wind circulation in the Salzach valley and the derivation of six characteristic phases of valley wind circulation, based on a thorough analysis of diurnal wind variations. In addition, there are a series of observations on mass transport of the valley wind, its vertical motion, and local peculiarities of the air flow in a canyon, etc.

Overview of Geography and Measuring Techniques

The Salzach Valley changes its orientation from WE to SN at the village of Pongau and then becomes narrower, assuming an almost canyon-like character as it forces its way between the limestone masses of the Hagen Mountains to the west and the Tennen Mountains to the east (see Figure 1 and Figure 7). Here, next to the narrow river bed, there is barely room for the highway and the double tracks of the railway between Salzburg and Innsbruck. The site of the single theodolite radiosonde station was immediately south of the narrowest part of the valley, on the east bank of the Salzach River next to the Sulzau train station.

Five km to the north of this site, the valley is almost totally closed by Ofenauer Mountain, approximately 900 m high and part of the Hagen Mountains [Ed. Note: To the west]. The river forces its way out of the mountains through the well-known gorge called "Salzachoefen". At this point, the road detours to cross the 562 m high Lueg Pass, and the train tunnels under Ofenauer Mountain. After the Salzach breaks through this constriction, the valley widens immediately into a basin, approximately 3 km wide, which then opens further to the Salzburg Basin, where the river enters the Alpine foothills. The second pibal station was established at Russeck, 4 km north of Lueg Pass, between Golling and Kuchl. The theodolite baseline at Russeck was over 1 km long.

The third site, Lehngries, lies in the Lammer Valley, 5 km ESE of the village of

Golling. The Lammer River, a tributary to the Salzach, flows around the east and north flanks of the Tennen Mountains (to the east of the Salzach) and flows into the Salzach south of Golling. A theodolite baseline of barely 500 m was established with difficulty in the rather narrow valley. A reasonably good view was available from the end points of the baseline.

Thanks to the efforts and the willingness of the former head of the Salzburg Airport Weather Service, Mr. Wagner, Weather Service Inspector, several pibals were launched from the Salzburg airport during the maneuver period for comparative purposes.

With this distribution of measurement stations it was possible to study some hitherto unanswered questions about valley winds. The following is a partial list of information obtained:

- simultaneous measurements of valley winds at three points in a longitudinal section of a valley (Salzburg, Russeck, Sulzau);
- the nozzle effect of a valley constriction (Lueg Pass);
- divergence of flow at the fork of two valleys (Salzach Valley and Lammer Valley);
- vertical component of the valley wind;
- the first aerological sounding of the atmosphere within the mountains by means of a series of radiosonde ascents (Sulzau) and comparison with the corresponding relationships over the plains (Munich);
- and more.

In Figure 6, the topographic cross sections of the valleys through the observation sites provide more detailed information on the terrain at the sounding sites. The cross sections show the different shapes and sizes of the valleys. The valley at Russeck is flanked on the west by the Hoher Goell (2519 m), a steep limestone massif that separates the Salzach Valley from the basin of the Koenigs Lake. The gently climbing ridges opposite the Hoher Goell, covered with meadows and pastureland, belong to the part of the Salzburg Alps which serves as a transition to the less rugged terrain of the Salzkammergut. The Salzburg Alps culminate here at about 1700 m with a few rounded peaks.

The valley landscape is quite different at the Lueg Pass near Sulzau. Instead of a U-shaped profile as at Russeck, the valley here is V-shaped. The slopes of the Tennen Mountains and the Hagen Mountains rise steeply on both sides. The slopes are bare, only sparsely forested with dwarf mountain pines and isolated conifers. On both sides of the valley the mountains form extended Karstic limestone plateaus, which rise on the average 1600 m (maximally 1900 m) above the valley.

The cross section at Lehngries is similar. Here, too, the valley is clearly V-shaped though not quite so precipitous. It is bounded on the left by the north wall of the Tennen Mountains and on the right by northern spurs of the Salzburg Limestone Alps (Schwarzer Mountain 1585 m). The valley floor is nowhere more than 500 m wide.

Salzburg lies in the broad basin where the Saalach River flows into the Salzach River. It is surrounded by the Salzburg and Berchtesgaden mountains in a wide semicircle to the south and by the Bavarian Plateau to the north. It can be considered a valley station only in the broadest sense of the term.

The difference in altitude of the three stations is minimal. Russeck lies at 470 m above sea level; Sulzau and Lehngries at 510 m. On the other hand, the areas of the three valley cross sections differ considerably. They are roughly 11.0 km^2 at Russeck, 5.2 km^2 at Sulzau, and 3.3 km^2 at Lehngries.

The actual series of soundings at these three stations began on September 4, 1941 at 18:30 DSZ [Ed. Note: German Summer Time] and ended after almost two complete days (41 hours) with the 11:00 DSZ ascent on September 6, 1941. In all, close to 50 pibal ascents were made at each station (46 at Russeck and Sulzau and 50 at Lehngries). Thirteen of the soundings at Sulzau carried radiosondes, unfortunately without instruments to measure humidity. All but two could be tracked. Salzburg (434 m) participated with a total of 7 pibal releases (single theodolite tracking) from 13:30 DSZ on September 5 to 06:00 DSZ on September 6. The weather was excellent during the measurement period. At midday on September 4, an occluded front, which had become inactive, passed through and ended a period of bad weather. The research area then lay on the SE edge of a high pressure system over central Europe, which developed a small, independent core of high pressure over the research area on September 5.

This series of soundings was preceded on September 3 by test ascents, which were made at three locations in the narrow valley south of Lueg Pass. Single theodolite tracking was used. These ascents were made to find the site in this area with the best natural visibility. In addition to meeting their intended purpose, the test ascents also produced some noteworthy information on local flows in a valley, even though the weather conditions were less favorable at that time.

Because of the restricted visibility in the curved valleys at Sulzau and at Lehngries, it was advantageous to work with ascent speeds of 200 m/min for the pibals and 400 m/min for the radiosonde balloons. Readings were taken at 15 second intervals. In the broad valley at Russeck, ascents were made at 100 m/min, at least at night, so that readings could then be reduced to every half minute. Most of the pibal ascents were cut off after about 15 minutes, if indeed visibility permitted tracking for that long. (At night ground fog occasionally occurred at several places.) There were two reasons for this 15 minute cut off: first, the interesting part of the circulation was observed in that time. Second, ascents had to be made at 1/2 hour intervals during the times of the actual artillery exercises. The graphical analysis was done immediately after the measurements were taken (sometimes by lantern light in improvised shelters in the field) by the trained staff of the weather observers section. It should be noted that all times used in the following are given in DSZ, which is 1 hour before Middle European Time.

Air Flows Described by Isopleths, Vertical Motions

The best overview of the temporal and spatial development of the local wind systems, which flow mainly parallel to the axis of the valley, is provided by graphs, with time as the abscissa and height above the ground as the ordinate. Lines of equal velocity of the valley wind component are then drawn on this graph. These isopleth diagrams, which ignore the unimportant and uninteresting cross-valley components, reduce the whole question to scalar units. Such diagrams have already stood the test in early investigations of this kind and are used here again, making comparisons with earlier results easier (Figure 2).

Because the three main sites are at practically the same elevation above sea

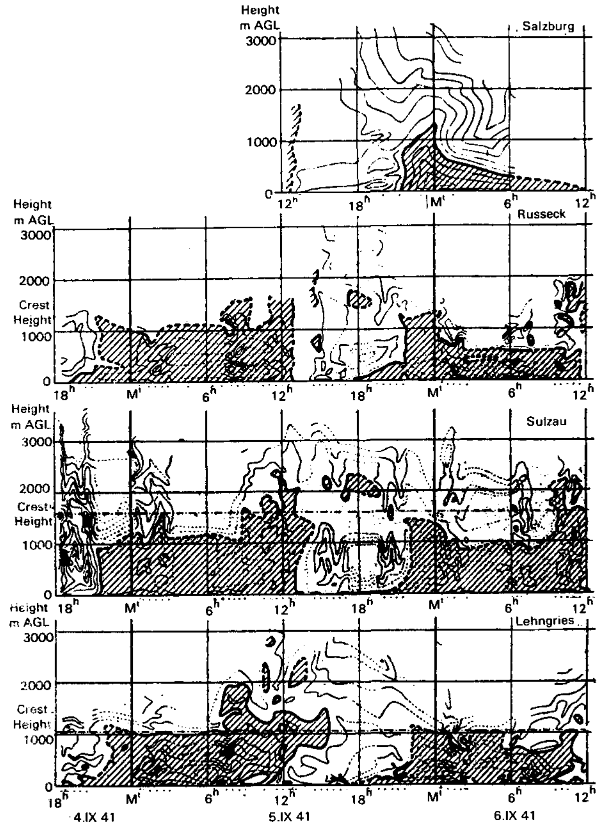

Figure 2. Velocity isopleths of the valley wind. Intervals: 2 m/s. Heavy lines = 0 m/s. Dashed lines = interpolated values. Hatched area = down-valley wind regime. Dots along the lower margin of each diagram indicate ascent times.

level (approximately 500 m), the elevation of the local wind regimes can be compared directly. Only at the auxiliary site in Salzburg is it important to remember that the starting point is approximately 70 m lower. In the diagrams, areas with a down-valley wind component are hatched to distinguish them from those with an up-valley wind component. Isopleth intervals are 2 m/s. Times of the soundings are indicated by dots at the lower edge of the figure. Sections of curves that have been interpolated are indicated by dashed lines.

A strong similarity in the patterns for the four stations is apparent at first glance. The similarity is the weakest at Salzburg, but it should not be forgotten that a considerably smaller number of observations serves as a basis for this diagram. For this reason, too, the curves look smoother than for the other stations. They are also smoother than might be expected in reality, when one considers the open location of Salzburg. Compared to earlier analogous diagrams, for example the first ones of Innsbruck, the relationships represented here give a much more balanced impression. This is especially true for Sulzau, which, for this reason as well as others, is particularly well suited for a detailed analysis.

Let us consider first the main characteristics common to all four diagrams. The flow at all stations shows a simple diurnal cycle within the first hectometers above the ground. Down-valley winds predominate during the first half of the day. At noon, these are replaced by up-valley winds, which last until about two hours before midnight. Up-valley winds reach their maximum rather late, at or shortly after 18:00. There are two maxima for the down-valley wind, corresponding to the effect of the semidiurnal pressure wave as described by Wagner [1, p. 341]. One occurs soon after the beginning of the period of down-valley flow (at or shortly after midnight). The second, usually stronger, maximum occurs toward the end (late morning or forenoon). A double stratification was also recognized spatially and will be discussed later.

In cases studied before, the up-valley winds in main valleys and side valleys were always stronger than the down-valley winds, averaged over the whole valley wind layer. Just the opposite was observed at Salzburg and Lehngries. Here, the down-valley wind was the stronger flow, in spite of the fact that during the entire time the gradient wind,[a] at least at higher levels, had a

(a) Gradient wind and geostrophic wind are used interchangeably here.

component parallel to the up-valley wind. For the time being it cannot be decided if this is the rule or an exception for these sites. It is possible that the season, late in the year, with its predominant radiation loss plays a role here. In addition, one must consider Salzburg's location at the opening to the plains, where, according to Hann, the down-valley wind is usually especially strong.

The down-valley wind period was observed twice in its entirety during the course of the field experiment, except at Salzburg. The up-valley wind period was measured completely only on September 5. On September 4, since measurements were not begun until after 18:00, the up-valley wind was measured only as it weakened. This day also presents a relatively disturbed picture. The development of the up-valley wind seems to be unusually strong, at least in the narrow section of the valley at Sulzau. Here, up-valley wind components over 14 m/s are indicated in the layer up to the crest altitude. The superposition of the gradient wind is, without a doubt, responsible for this. At the time, the gradient wind had approximately the same orientation as the up-valley wind and was locally enhanced by the nozzle effect in the canyon-like section of the valley. On the evening of September 4, irregularities in the thermal stratification appear at higher levels. This will be discussed later.

On the following days, the upper level winds became more westerly [Ed. Note: Shifted from the north]. The upper boundary of the up-valley wind was then distinguished by a clear minimum of the wind speed, just as required by Wagner's theory [2, p. 422 ff]. On both nights, the corresponding boundary of the down-valley wind is marked directly by the zero line of the isopleths. Above a thin, calm layer, the wind shifts directly into a flow with the opposite component.

The vertical extent of the local winds decreases everywhere on the second night, probably as a result of changed weather conditions. This is clearest in the open part of the Salzach Valley. At Russeck the drop was 500 m (from 1100 to 600 m a.g.l.). The experiences at Innsbruck showed that the conditions for the development of the valley wind are particularly favorable immediately after a period of bad weather. This is especially true after a cold front passage. In the fresh, clean, cold air the local pressure difference between valley and plains, caused by radiation and heating, increases. In our case, moreover, the temperature contrast during the first night between the air already warming over the plains and the cold air still in the mountains would probably have been greater than normal.

Even if this height difference of the upper boundary of the down-valley wind on the two experimental days is disregarded, the vertical extent of the valley wind is greater during the day than at night. This is also true in all valleys studied previously and is explained by the scheme of valley wind circulation developed by Wagner. During the day, the thermal pressure gradient between the plains and the mountains, which causes the valley wind, certainly has a greater vertical extent than at night. This vertical extent is further enhanced by the convective mixing of the valley air and by the dynamic warming of the return flow of the lateral slope flow circulations, as it sinks over the middle of the valley. At night, when the air generally is more stable, this effect is intensified in the valleys, due to the influx of cold air from all sides. Above this cold pool, warming due to sinking is observed.

Perhaps this also explains why on the morning of September 5, a pronounced bulge appears in the zero isopleth at all three stations in the final phase of the down-valley wind regime. With the heating process beginning, thicker and thicker layers become involved in the air exchange between the mountains and the plains. The temperature gradient has not yet reversed and still points from the plains toward the mountains. Once again, the down-valley wind increases in height before it vacates the field for the up-valley wind. A supplementary analysis of the first valley wind investigation at Innsbruck [3] revealed in the isopleth diagrams an analogous increase in the vertical extent of the down-valley wind in the morning hours. (Compare, for example, the isopleths of June 19, and July 12, 1929).

This investigation cannot answer the question of a possible time delay in the onset of the valley wind along the length of the valley. To do so, a significant increase in the number of ascents at the times of the daily wind reversal would have been necessary. This was not possible for external reasons. However, the available observations seem to support Wagner's opinion that the onset of the valley wind takes place simultaneously in the entire valley, as long as the very lowest layers are disregarded, i.e., only the regime of the "proper" valley wind is considered, and as long as the valley floor is not steep. The only case in which simultaneous measurements reveal a distinct delay in the onset of the up-valley wind as one proceeds into the valley was in the Kloster Valley on the west side of the Arlberg Pass [4]. However, the valley floor climbs steeply between the two measurement stations, Stuben and St. Christoph.

The valley wind begins earlier in the layers near the ground than in those above. The time difference varies from station to station and can be greater than 4 hours. This was especially noticeable at Sulzau, where a significant flow down the valley began at 16:30 as soon as the sun disappeared behind the ridges of the Hagen Mountains. At the sa time, the main up-valley wind above these lower layers had just reached its full development. The onset of a down-valley flow near the ground was accompanied by a drop in the temperature near the ground. On September 5, the temperature dropped 10°C within 30 minutes and 7°C within 15 minutes. See Figure 14 in Part II. [Ed. Note: The second part to this paper was never published. The Ekhart translation which follows this paper apparently contains the information he had intended for Part II.] Without a doubt, this is the premature appearance of the cold air drainage, which is especially clearly developed here due to special orographic relationships. Earlier I gave this drainage the name "downslope wind along the valley floor" to indicate its genetic relationship to Wenger's slope winds. At Russeck and Lehngries, the cold air drainage near the ground extends considerably higher up, apparently fueled by the especially well developed lateral slope winds here. During the night it probably increases to as much as 400 m, as shown by the very pronounced double structure in the isopleth diagrams as well as in the mean vertical velocity profile of the down-valley wind (Figure 5). This is not seen in the relevant profiles at Sulzau because the minimum speed between the two components of the down-valley wind is probably within the lowest 50 m interval between observations. The reason the nocturnal cold air flow along the valley floor is not stronger here is probably that the lateral slope winds, the main source of nourishment for the cold air drainage, are only minimally developed.

Some support for this assumption comes from the diurnal variation of the observed ascent speeds of the balloons at the pibal stations. Based on our general knowledge, higher ascent speeds would be expected during the day than at night due to the increase in convection and turbulence. However, as the observations have shown, this is true only at Sulzau, where the up-valley wind soundings in the layer from the ground to 1000 m a.g.l. show an ascent speed on the average about 0.2 m/s greater than for the down-valley wind soundings. (Sulzau was equipped with radiosondes, and therefore the height of the soundings could be measured precisely.) In contrast, at Russeck, in the broad Salzach Valley with its more gently sloping sidewalls, the balloons rose on the average about 0.4 m/s slower in the first 1000 m with up-valley winds than with down-valley winds. In the Lammer Valley, at Lehngries, the balloons rose on the average 0.3 m/s slower in the first 700 m.

A rather informal explanation for these differences could be the following: the slope winds can develop in a completely normal fashion on the slopes at Russeck and Lehngries because the slopes are gentler and the valley is broader. The more favorable these conditions are (at Russeck more so than at Lehngries), the better the development of the slope winds. As a result of the return flow of the slope wind cross circulation over the middle of the valley, the valley wind flow shows a tendency to descend during the day (upslope winds) and a tendency to ascend at night (downslope winds). The valley cross section at Sulzau lacks sufficient space for the lateral slope winds due to the steepness of the sidewalls on both sides and the narrowness of the valley. Therefore, the valley wind itself has to support the vertical transport necessary for closing the circulation. Thus, the up-valley wind exhibits a small but usually measurable upward component, and the down-valley wind exhibits a downward component. The cross section scheme of valley winds in the form drawn by Wagner [see 2, Figure 4, p. 439] apparently is not valid for such narrow canyons as the Salzach Valley near Sulzau. The slope winds necessary for the maintenance of the valley wind are missing. Nevertheless, the development of the valley wind can proceed undisturbed even here because the flow, which overcomes the constriction solely because of its inertia, meets with "normal" conditions again above and below the constriction, i.e., a broader U-shaped valley and more gently rising slopes covered with vegetation.

Figure 3 presents the daily variation of the ascent velocity of the pibals at Lehngries for the lowest 700 m calculated as deviations from the mean value (211 m/min). Two "anomalies" should be mentioned, which, exactly twelve hours apart, disturb the otherwise quite smooth curve. First, at about 06:00, a distinct dip occurs within the period of above normal ascent velocities, i.e., with positive vertical components. At 18:00 a disturbance occurs in the form of higher than normal values during the period of below normal buoyancy (i.e., downward vertical motion of the air). Evidently these are the reversal times of the lateral slope winds, which are known to develop more or less ahead of the valley wind. (At 06:00 the transition is from downslope to upslope wind; at 18:00 the reverse.) As Wagner assumed for thermodynamic reasons

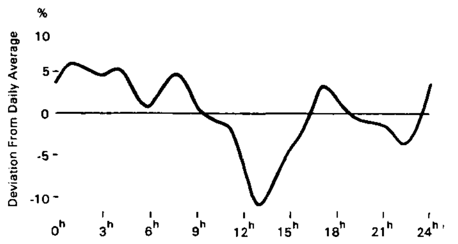

Figure 3. Diurnal variation of the ascent velocity for pibals at Lehngries as deviation from daily means (September 4 - 6, 1941).

[see 2, p. 441 ff], the valley air as a whole sinks in the morning and rises in the evening.

Mean Balloon Paths

Diagrams of the mean flight paths (projected horizontally) of the balloons in up-valley winds and down-valley winds provide an overview of the main characteristics of the valley wind circulation. These characteristics are clear, even though there are real, and sometimes random, temporal fluctuations of speed. Figure 4 gives the mean flight paths for each of the four stations. To construct the diagrams in Figure 4, all ascents at a station during the up-valley period, and then all ascents during the down-valley period, were averaged vectorially. The short lines across the curves indicate 100 m height intervals (relative heights). The length scale is the same for all curves, but the velocity scale varies according to the different average ascent velocities of the balloons at each station. North is at the top of the figure. The down-valley direction is indicated by a dashed arrow for the pair of curves for each station.

Here it becomes immediately apparent that the paths of the balloons closely follow the main axis of the valleys from the ground to just below crest altitude. This is also true qualitatively for individual cases. The lateral displacements within the valley wind regime are minimal

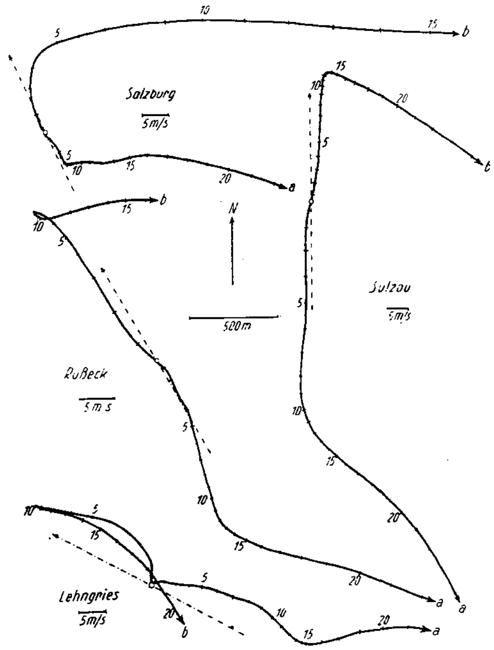

Figure 4. Mean pibal flight paths during the up-valley wind period (a) and the down-valley wind period (b). Numbers indicate hectometers above ground.

for each individual ascent. The changing interval of the height marks demonstrates the variability of the flow velocity with height. The velocity first increases from the ground up and reaches a maximum a few hectometers above the valley floor. It then decreases steadily to a minimum at crest altitude. This coincides with a very sharp bend in the path. In some cases, in particular for down-valley winds at Lehngries, this bend degenerates to a 180° change in the wind direction. Therefore, the valley wind regime extends, on the average, up to this level. What lies above that is determined mainly by the gradient flow, which increases in strength as the height increases. In our case the gradient flow is usually from the W-NW. It is not surprising that the strength and direction of the gradient flow vary for the up-valley and down-valley winds, since changes in the general pressure distribution caused by weather changes occurred during the course of the experimental period.

However, one would expect that for comparable curves, i.e., for up-valley winds and separately for down-valley winds, the upper air flow at the different sites would at least have the same direction since the curves are calculated for an almost identical period of time for all stations except Salzburg. It is easy to understand, however, why this is not the case. First, the method of difference applied here, vectorially and individually for each station, does not completely compensate for the loss of observations, which increases with height and is different at each station. Secondly, real differences certainly play a part. There is, for one thing, the addition effect of the upper compensating flow of the valley wind circulation. Most important, however, is the orographic influence on the flow, which is different at each site. The varying course of the flight paths in the geostrophic wind layer shows clearly that the orientation of the valley influences the direction of the flow, even at heights above the ridges. The influence is greater, the narrower the valley. In this way the gradient wind, blowing essentially from the west above the open Salzburg basin, is redirected into a WNW-NW flow above Sulzau - especially during the day (up-valley wind!)[a]. This directional influence of the lower "channelized" flow on the upper "free" flow is apparently accomplished by the exchange between the air closed in the valley and the free atmosphere above the mountains. (See also Part II).

[a]To counter the objection that the change in direction of the gradient wind between Sulzau and Salzburg resulted from changes over time, the up-valley wind pilot balloons at Salzau on September 5, i.e., the experimental day at Salzburg, were also compiled for a mean flight path. The results were essentially the same.

Mean Velocity Distribution with Height. Longitudinal Section Scheme of the Valley Wind System

The above figures of the mean flight paths show satisfactorily that we can limit ourselves to the valley component of the flow when deriving the most important characteristics of the valley wind. This has the advantage of allowing scalar analysis. The upper boundary of the valley wind, its mean and maximum velocity, as well as the height of the velocity maximum under quasi-stationary conditions can best be read from vertical profiles of the distribution of the valley component of the wind. These profiles are obtained by averaging all ascents during up-valley wind periods and during down-valley wind periods. The drawing in Figure 5 is cut off at the upper boundary of the valley wind for the sake of clarity.

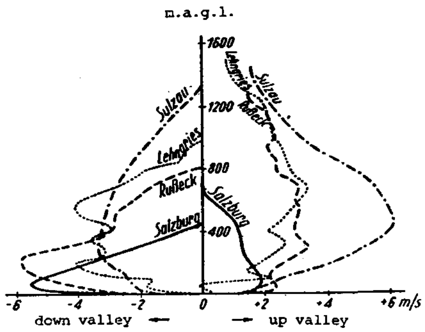

Figure 5. Average distribution of velocity with height for the valley winds.

The following summary table results from Figure 5.

Structure of the valley wind

Station	altitude (m a.s.l.)	upper boundary (m a.g.l.)	horizontal velocity (m/s): mean value of the whole layer	horizontal velocity (m/s): maximum	horizontal velocity (m/s): height of maximum (m a.g.l.)
			Up-Valley Wind		
Salzburg	434	660	1.14	1.9	80
Russeck	470	1270	2.50	3.4	300
Sulzau	510	1460	3.91	6.2	500
Lehngries	510	1480	2.10	3.3	680
			Down-valley wind		
Salzburg	434	450	3.00	5.5	70
Russeck	470	820	3.17	1.58 2.28	230 480
Sulzau	510	1360	2.16	3.3	350
Lehngries	510	970	2.56	1.41 2.41	200 560

First we should mention that the double layering of the down-valley wind discussed above leads to a splitting of the velocity maximum into two separate maxima. However, this is apparent only at Russeck and Lehngries. At Salzburg and Sulzau the resolution of the observations is insufficient for this purpose.

We can clarify the remaining content of the table by means of the following diagrams. The left side of Figure 6 shows a longitudinal section of the Salzach Valley with the velocity isopleths of the up-valley wind (above) and of the down-valley wind (below). These were calculated from the mean values of the along-valley component of the wind at the different pibal stations and at different heights and therefore represent the quasi-stationary state of motion. The right half of the figure shows the vertical extent of the valley wind system by means of four superposed cross sections of the Salzach Valley and one cross section of the Lammer Valley.

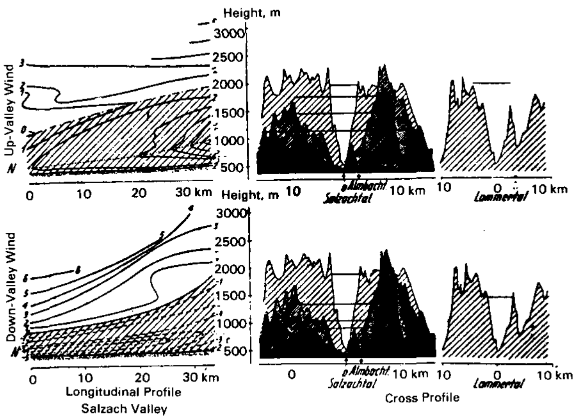

Figure 6. Scheme of a longitudinal section of the valley wind in the Salzach valley. The altitude scale has been increased by a factor of 10. Left: velocity isopleths of the valley component. Heavy lines: up-valley wind component; dashed lines: down-valley wind component; valley wind region hatched. Right: Cross section through the Salzach Valley and the Lammer Valley with upper boundary of the valley wind. The shading of the profiles lightens in the up-valley direction. The superposed profiles are for the following locations: Salzburg airport, Kaltenhausen (near Hallein), Russeck, and Sulzau. The profile of the Lammer Valley at Lehngries is on the right.

(The shading is lighter for each cross section further up the valley). The altitude scale has been increased by a factor of 10.

This figure is the first attempt to describe in the proper order of magnitude the spatial structure of the valley wind in the longitudinal section of a valley. Of course, similar attempts were made earlier with the glacier wind and the Maloja wind of the Engadine [5, 6]. In the case of the Maloja wind, the attempt was based on only a few measurements taken at different sites in the valley and at different times. In the case of the glacier wind, a schematic picture was intended to clarify data, part of which had been obtained indirectly. However, this had to be limited to a qualitative form due to the lack of sufficient data. The present diagram is completely new for the valley wind, because up to now there have not been simultaneous measurements at several points in the same valley. This not only makes it possible to place a longitudinal section scheme along side Wagner's cross-sectional scheme of the valley wind, but it also makes it possible to draw quantitative conclusions, as we will see below.

The valley wind regime is clearly separated from the upper air winds during the up-valley wind period and during the down-valley wind period. In the case of down-valley winds, they are separated directly by the zero velocity line. In the case of up-valley winds, they are separated by a distinct zone of minimum velocity values between the valley component of the lower flow and the valley component of the upper flow, both of which have the same sign.

The regime of the down-valley wind is wedge-shaped, decreasing sharply in depth from the mountains toward the mouth of the valley. At the same time the velocity in the direction of flow, i.e., out of the valley, increases in accordance with Hann's statement, which was mentioned above. The height of the velocity maximum within the down-valley flow also drops with the upper boundary, although to a lesser extent. Correspondingly, the gradient wind reaches deeper and deeper as it flows toward the mouth of the valley, so that considerable differences in velocity occur at the same level over relatively short distances along the valley. For example, at 2000 m, the transition from down-valley wind to a gradient wind blowing up the valley is taking place in the mountains (Sulzau), while wind speeds of 6 m/s toward the mountains are being measured over the Salzburg basin.

The up-valley wind regime, in contrast, increases in height in the direction of the flow, thus also forming a wedge which decreases in depth toward the mouth of the valley. At the same time it also increases in intensity. Over Salzburg the up-valley wind attains a mean maximum velocity of barely 2 m/s at 500 m. At Sulzau, the velocity is already 6 m/s at a level approximately 500 m higher. On the other hand, the velocity of the gradient wind levels out horizontally at a lower height than in the case of the down-valley wind. This can be seen in the 3 m/s isopleth, which runs horizontally at a height of 2350 m along the entire section.

Thus the regimes of the up-valley wind and the down-valley wind have the same shape in the longitudinal section: in both cases it is a wedge with the point towards the mouth of the valley. In a purely qualitative sense this also describes the ratios of valley cross sections, as can be seen immediately in the profiles on the right side of Figure 6. The lateral ridges increase in height in the up-valley direction with the valley floor running almost horizontal. According to Wagner's empirical findings, the diurnal temperature range decreases more slowly with height in the valley floor-crest altitude region than over the plains. In other words, the daily range of the average temperature of the valley air increases steadily in the up-valley direction. Along with this, the thermally caused pressure gradient between the plains and the mountains during the day and between the mountains and the plains at night involves thicker and thicker layers as one goes farther into the valley. This is true as long as the elevation difference between the valley floor and the lateral ridges continues to increase. In our case, therefore, the upper boundary of the valley wind should continue to increase beyond Sulzau, the southernmost pibal station, and should presumably culminate at the big bend of the Salzach River at Pongau, where the Hoch Koenig massif, almost 3000 m high, rises on the left side of the river opposite the Hohe Tauern on the right side. Although the lateral ridges increase in height, the width of the Salzach Valley decreases as one goes farther into the Alps. The question therefore arises to what extent the variation in the flow velocity of the valley wind corresponds to the change of the cross section of the flow, taking into consideration the air density, which varies with height. Expressed another way, is the continuity condition met for the valley wind? We therefore calculate - neglecting possible horizontal density variations - the total mass flux in the up-valley and down-valley flow for the quasi-stationary situation represented in Figure 5 using the mean vertical profiles of velocity. The result is as follows.

Let us first consider the down-valley wind. Under quasi-stationary conditions 8830 tons of air per second flow northward with the down-valley wind through the canyon at Sulzau, our southernmost station in the Salzach Valley. North of Lueg Pass (see Figure 1) this air flow is augmented by the down-valley wind from the Lammer Valley. According to the observations at Lehngries, this amounts to 7870 tons per second on the average, i.e., only 10% less than the down-valley wind in the Salzach. The main flow, thus augmented, continues past Russeck, for which a mass flux of 16,250 t/s was calculated for down-valley wind conditions. The difference between this and the sum of the two partial flows just described is less than 3%. Farther down the valley we still have the Salzburg station where the valley opens onto the plain. Here, under down-valley wind conditions, a flow of 25,450 tons of air per second was calculated. This is half again (9200 t/s) more than at Russeck. There must, therefore, be an additional

Mean flux of air mass of the valley wind circulation, t/s

	Up-valley wind	Down-valley wind
Salzburg	14710	25450
Russeck	31350	16250
Sulzau	18310	8830
Lenhgries	13210	7870
Sulzau + Lehngries	31520 = Russeck	16700 = Russeck
Salzburg - Russeck	-16640 = ?	9200 = 4050 Almbach Valley + 5150 Saalach Valley

contributing flow between Russeck and Salzburg. A glance at the map shows the source: one part is from the Almbach Valley, a tributary of the Salzach Valley on the west side which drains the Koenigs Lake and flows into the Salzach about 10 km above Salzburg. Another part comes from the Saalach Valley, which enters the Salzburg basin from the SW. In order to explain the surplus of mass at Salzburg, we need a rough calculation of approximately how much each of these two valleys contributes.

For this purpose we draw a terrain profile across the Salzach Valley and the Almbach Valley about halfway between Salzburg and Golling, i.e., where the two valleys are almost parallel and are separated only by a low ridge, the Goetschen (930 m). (This is in the area of Hallein, the second profile from the front on the right hand side of Figure 6.) This separating ridge lies below the upper boundary of the Salzach down-valley wind, as can be seen clearly in the profiles on the right in Figure 6. The Salzach down-valley wind, therefore, flows over the ridge and merges with the down-valley flow of the Almbach Valley on the other side. The two down-valley winds thus form a single flow here. This is analogous to the situation described by Wagner [2, p. 420] in which the valley winds of the Etsch Valley and the Sarca Valley merge above the dividing wall of the Monte Baldo to form a unified valley wind. If we now measure the cross section of the Almbach Valley and assign the same vertical distribution of the down-valley wind to it as exists in the Salzach Valley at the same width, which can be easily derived by interpolation from our longitudinal profile on the left side of Figure 6, we get a flow of 4050 t/s, which feeds the Salzach down-valley wind. Accordingly, the remaining 5150 t/s must be brought into the Salzburg basin by the down-valley flow from the Saalach Valley. That this rather large valley delivers only about 25% more than the smaller Almbach Valley is explained by the orography of the Saalach Valley. The streambed of the Saalach down-valley wind already has a direct exit to the plains between Lofer and Reichenhall, above the point where it enters the Salzburg Basin. This exit is the low pass between the Rauschberg and the Hochstaufen into the valley of the Rote Traun, which drains to the north.

The relationships are similar under up-valley winds, which throughout carry about double the mass of the down-valley winds. The up-valley wind over Russeck forks south of Golling. About 2/5 branches off into the side valley of the Lammer, and the rest (58%) continues to flow in the main valley. This is shown clearly by the values for the three pertinent stations (Russeck, Lehngries, Sulzau) given in the table above. The mass flux calculated for Salzburg, which doesn't even amount to half (47%) of that at Russeck, at first seems to be surprisingly small in comparison. The flow at Salzburg should exceed that at Russeck, since the air transport up the valley from Salzburg must also feed the up-valley winds in the Saalach Valley and the Almbach Valley. This apparent discrepancy can, however, be explained. In calculating the flow over Salzburg, only the region between the Hohenstaufen (the outermost peak on the right edge of the diagram in Figure 6) and the Heuberg (the broad hill on the left in the figure) defined the profile of the "valley" and therefore of the up-valley flow in the first approximation. However, the upper boundary of the up-valley wind in the Salzach Valley is obviously above the Heuberg, so that air from the region to the left of the Heuberg (geographically east) is also drawn into the flow of the up-valley wind. However, since there is no precise information about the extent of this drainage area, it doesn't make sense to correct the above figure for the mass flux of the up-valley wind at Salzburg. We will simply state that it is, without a doubt, too low.

Nevertheless, we can still give an essentially correct description. The air flows from all sides out of the northern foreland, collects in the funnel-like Salzburg Basin, and then flows as the up-valley wind through the Salzach Valley and its tributary valleys into the interior of the Alps. Conversely, the air carried by the down-valley wind to the mouth of the Salzach Valley spreads fan-like into the plains. This contracting and expanding has its thermal effect, as will be shown later (see Part II).

It cannot yet be concluded from the wedge shape of the valley wind or from the continuity of the mass flux within the valley wind flow that the valley wind air as a whole moves up or down, depending on whether the flow being considered is the up-valley wind or the down-valley wind. In other words, the consistency of the mass <u>flux</u> still does not justify the assumption that the air <u>mass</u> is also constant. At least this conclusion would not be convincing, since the upper surface of the valley wind is by no means rigid and impenetrable. On the contrary, air can very probably enter into the valley wind circulation from the regime of the geostrophic gradient wind, or exit out of the valley wind circulation into the gradient flow under down-valley wind conditions. But then, since mass continuity of the valley wind flow is realized to a large extent, just as much air must be removed from the valley wind regime. According to Wagner's concept, this is a function of the lateral slope winds, as is well known. It is, therefore, not only possible, but very probable that the valley wind flows essentially horizontally, in spite of its variable vertical extent in the direction of the flow, as long as a steep slope of the valley floor doesn't force the lowest lines of the flow (and only the lowest) to ascend or descend.[a] This therefore satisfies the thermodynamic conditions of Wagner's theory.

<u>Details of the Down-valley Wind at the Lueg Pass (see Figure 7)</u>

In connection with this general picture of the flow stratifications of the valley wind in the longitudinal section of the Salzach Valley, a few detailed observations, made during the preliminary test studies on September 3 in the Lueg Pass area, should be added. At that time,

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[a] A few words concerning valley wind terminology may be appropriate here. Wagner's theoretical studies have made it necessary to define the genetically based differentiation of diurnal mountain winds into valley and slope winds. Wagner [7, p. 227] suggested the terms "up-valley and down-valley winds" and "upslope and downslope winds" to replace the old "mountain and valley winds" which included both valley winds and slope winds. This new terminology is not very successful either, as far as the valley winds are concerned. It could all too easily be misinterpreted, and already has been, as if it described a genuine "up wind and down wind." Wagner himself established and repeatedly supported the concept, confirmed by experience, of a valley wind flowing horizontally. According to the meaning of the Wagner terminology, the adverbs "up" and "down" do not refer to the noun "wind", but to "valley": up-valley wind = a wind in the direction of the climbing valley (likewise down-valley wind = a wind flowing toward the mouth of the valley). Perhaps these misunderstandings could be avoided if one spoke of "in valley winds" and "out valley winds". [Ed. Note: Research teams in Austria prefer the terms "in valley wind" (Taleinwind) and "out valley wind" (Talauswind). German scientists prefer the terms "valley wind" (Talwind) and "mountain wind" (Bergwind)].

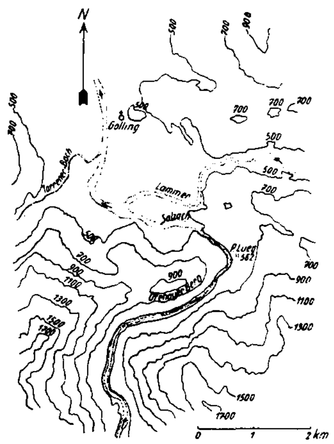

Figure 7. Orography of the Lueg Pass area.

single theodolite pibal ascents were made from three sites, about 1 1/2 km apart, in the narrow part of the valley at Sulzau. The middle of the three stations was later used as the ascent site at the Sulzau train station. The five sets of simultaneous trackings were made during a two-hour period (from 9:00 to 11:00) at half hour intervals. The weather was dull and cloudy, it had rained heavily on the preceding days, and it had snowed in the mountains down to 1800 m. During the measurement period, the sky was almost totally covered with Ac and As clouds drifting in from the north, forming some fracti along the peaks and ridges.

Driving the car over the Lueg Pass toward the south, a rather strong down-valley wind was observed immediately below the top of the pass. A little farther up the slopes a gusty wind, coming out of the canyon, could be observed in the tree tops. In contrast, scarcely any air movement was perceptible at the ground farther up the valley, as had been the case to the north of the pass. There was just an occasional gentle breeze out of the valley (<1 Beaufort). At the sites themselves mostly calm air was noted. A down-valley surface wind of less than 1 m/s was measured with a hand anemometer only twice, at the northernmost of the three stations at the beginning of the measurement series and at the southernmost at the end.

This local increase of the wind speed right at Lueg Pass, and usually farther into the narrow part of the valley as well, is routinely observable. In every case, a storm blows down the northern slope of Ofenauer Mountain in the morning and forenoon hours with all the characteristics of a genuine katabatic wind. A little farther to the north, toward the community of Golling, it is already considerably weakened. In the winter, even with a heavy snow cover, the fields between the Ofenauer Tunnel and the railway bridge over the Lammer and the Salzach are often swept completely clear, and the snow is deposited in high drifts on the banks of the two rivers.

An especially impressive orographic cloud regularly forms over the top of Ofenauer Mountain as a result of the down-valley wind, even under otherwise cloudless conditions. This cloud demonstrates *ad oculos* the flow over an obstacle in the classic manner. From a distance, it appears to be motionless, resting over the mountain, sometimes enveloping the summit in fog. Coming closer, especially approaching from the side at the elevation of the pass, the flow becomes clearly visible. There is an ascending motion on the windward side and a waterfall-like descending motion on the lee side. This is, therefore, an habitual local foehn region. It shows all the characteristics of a foehn region, including a small-scale "foehn wall", created by the blocking effect of Ofenauer Mountain in the flow of the Salzach down-valley wind, which carries moist air out of the canyon. O. Pollack commented on this phenomenon three decades earlier, albeit in a somewhat obscure spot in the literature [8].

Returning to the soundings of the experiment series mentioned, we find weak down-valley winds in a 600 - 1500 m thick layer from the ground up at all three stations. The wind did not exceed 3 m/s at any height on any of the readings. Above this, the wind shifted almost abruptly and with a considerable increase in speed to a northerly direction, just the opposite of the wind in the lower layer. The following stratification resulted from averaging all measurements:

200 m a.g.l.	S-E	0.6 m/s
400	S-W	1.3
600	SSW	0.6
800	NNE	1.6
1000	NNE-M	3.0
1200	N	3.6
1400	NNW-N	3.8
1600	NNW-N	5.9 (crest altitude)
1800	N	7.9
2000	N-W	9.1
2200	N-W	10.4
2400	N-W	14.6

The upper boundary of the down-valley wind was easily determined for each case. (Measurements were taken at 15-second intervals with an ascent speed of 200 m/min.) Here, too, the wind boundary showed a significant drop in the down-valley direction at each of 5 pibal ascent times, just as was seen in the main investigation in the Salzach Valley between Sulzau and Salzburg. The average drop for the 5 series of soundings was 240 m over a horizontal distance of 2850 m, which is equivalent to a slope of almost 5°. The slope was not uniform over the entire distance, but flattened out somewhat downstream in such a way (see Figure 8) that it would run just over the top of Ofenauer Mountain, if it were extrapolated beyond the northernmost station. It is

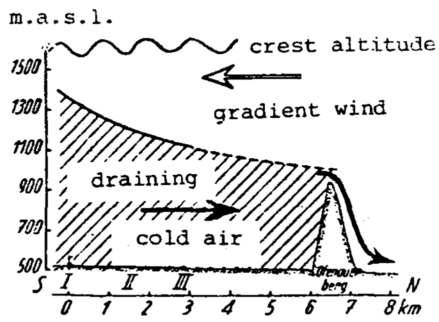

Figure 8. Upper boundary of the down-valley wind at Lueg Pass (Salzach Canyon) on the morning of September 3, 1944.

easy to understand why, with this vertical constriction of the cross section of the flow and with a horizontal constriction resulting from the narrowing of the valley, an increase in the wind speed is observed at Lueg Pass.

The observed shape and especially the steep slope of the upper boundary of the lower wind cannot be considered the norm for the down-valley wind in this part of the valley. Instead one has the impression that this was not so much the regular valley wind - indeed, the windspeeds are much too low for that - but rather the slow return flow of the cold air which had penetrated the Alps the day before, from the interior of the mountains to the plains. This was favored by the normal tendency to develop a down-valley wind at this time of day, even with cloudy weather.

With this drainage, the inclined upper surface of the cold air showed oscillations as it alternately flattened out and then became steep again. Perhaps it could be explained this way. A certain amount of cold air behind the barriers of the Tennen Mountains and the Hagen Mountains was necessary before the resulting accumulation of potential energy is sufficient to force the draining cold air through the nozzle-like constriction of Lueg Pass (see Figure 9). By this mechanism a pulsating "dripping" of the cold air, stored in the

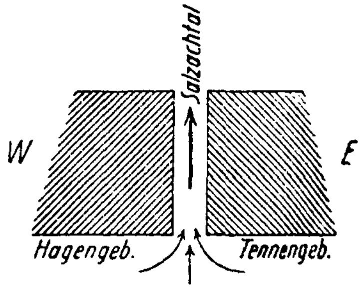

Figure 9. The nozzle effect in the canyon near Lueg Pass.

interior of the mountain range, results. This produces a flattening of the boundary surface with each "drip". In fact, the variations of the slope of the boundary surface (α) and those of the flow velocity (v) of the cold air are to a certain extent inversely proportional.

time (DSZ)	Δα (°)	Δv (m/s)
0900 - 0930	-0.5	+1.2
0930 - 1000	+2.3	-1.0
1000 - 1030	-1.5	-0.6
1030 - 1100	-5.3	+0.7

v = flow velocity for Station III averaged over the total down-valley layer

Of course, too much importance should not be placed on such isolated observations. However, one should remember that this type of pulsating drainage of cold air is not unknown in the mountains. Schmauss [9], for example, found such a drainage in the Partenkirchen basin before the foehn came through. Something similar was also seen in the case of glacier winds [5].

The purpose of this digression was to point out the favorable experimental conditions which the valley constriction of the Lueg Pass could offer for many problems of experimental flow research in nature.

Details of the Diurnal Wind Cycle Over Sulzau

Hodographs

We were interested in doing a thorough analysis of the soundings at Sulzau, especially the daily wind variation, for two reasons. First, the valley wind develops in a very typical fashion in the narrow part of the Salzach Valley. Second, the radiosonde ascents here also provide information on the thermal structure of the atmosphere.

In order to obtain a picture with the least possible influence from other phenomena, a diurnal cycle of the wind was derived from the available ascents by averaging over two extensive, intrinsically homogeneous layers, which show clearly the significant characteristics of concern here. The first layer, from the ground to 1000 m a.g.l., approximates the regime of the valley wind. The second layer, between 1700 m a.g.l. and 2500 m a.g.l. can be considered representative of the wind above the mean crest altitude. First, mean values of both wind components, valid at any given time for the entire thickness of the layer, were calculated by interpolation for each hour of the day, based on the entire series of measurements. From these mean values mean hourly wind vectors were compiled and put into a wind rose system, starting at the origin (+). The wind figures presented in Figure 10a are the result of smoothly connecting the endpoints of the individual hourly vectors. Shifting the coordinate system in the direction and in the amount of the daily resultant indicated by the arrow produces the corresponding system for the purely periodic cycle of winds. Arrows can be drawn on the wind figures from the new origin to the hourly marks to indicate the direction and magnitude of the hourly supplementary winds.

The curve of the lower layer is very simple and is at the same time a classic example of an undisturbed valley wind cycle. The entire diurnal cycle of air displacement is essentially reduced to a single forward and backward shift of the air parallel to the valley axis within 24 hours. The path is almost the same in both directions. The mean daily vector of the wind almost disappears, in that the ratios of the durations (τ) and the average strength (ν) of both flows are reciprocal: 9 hours (13:00 - 22:00) up-valley wind (↑) with 4.2 m/s on the average as opposed to 15 hours (22:00 - 13:00) down-valley wind (↓) with 2.5 m/s on the average. $\tau : \tau = \nu : \nu = 3:5$. However,

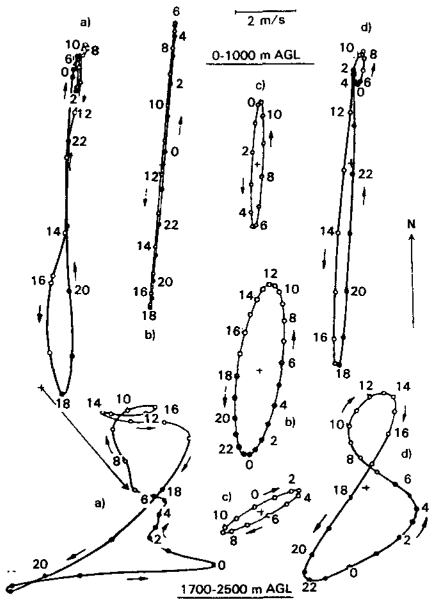

Figure 10. Daily cycle of winds over Sulzau, illustrated by vector curves (vectors pointing from the origin toward the points on the curve).

a. according to the observations
b. diurnal term of the harmonic analysis
c. semi-diurnal term of the harmonic analysis
d. sum of b and c

this amazingly complete compensation is probably just by chance. One can assume that here, as in most other valleys studied up to now, the up-valley wind is not only stronger, but also of longer duration than the down-valley wind in the middle of the summer. Maximally the up-valley flow reaches more than twice the strength of the down-valley wind (7.6 m/s as opposed to 3.4 m/s).

The daily cycle in the layer above the crest of the range (1700 to 2500 m) is entirely different. First, the influence of the gradient wind is seen in the presence of a mean daily resultant of more than 4 m/s out of the NW. In addition, the curve seems to be compressed longitudinally compared with the lower curve. It is, however, much broader. Thus, the shape of a double loop, which is also suggested in the lower wind figure, becomes clearly visible. The double loop divides the daily cycle into two separate halves, with the division coming almost exactly at 06:00 and 18:00. During the daytime hours, supplementary winds with a southerly component are dominant, moving clockwise. At night, supplementary winds with a northerly component are dominant, whereby the direction is counterclockwise. The wind reversal itself, with reference to the center of the figure (arrow point), takes place with the sun from 14:00 - 22:00 and against the sun the rest of the time.

At first glance a series of measurements made in less than two full days may not seem sufficient for a harmonic analysis of the diurnal wind cycle. However, the results of such an experiment show that quite useful and meaningful information can be obtained. Figure 10b and c present this analysis of the diurnal cycle in a full day and a half day elliptic oscillation. Figure 10d presents the vector curve for each of the two layers studied, obtained by graphical addition of the 10b and c curves.

The daily wind cycle within the valley wind regime (0-1000 m) is reproduced with impressive accuracy by the first two terms of the analysis. The strong predominance of the diurnal wave was expected. In addition, a conjecture by Wagner is confirmed by the semi-diurnal wave, which is also forced to develop only along the axis of the valley. In his "Theory and Observation of Periodic Mountain Winds," Wagner stated [2, p. 341]:

> The semi-diurnal variation in air pressure also has a certain influence on the asymmetric development of the mountain and valley wind. Since this semi-diurnal fluctuation is more strongly developed in mountain valleys than over the plains, the mountain wind is weakened at 04:00 (first maximum of the semi-diurnal wave). In contrast, at 16:00 the valley wind is intensified (second maximum).

In fact, the semi-diurnal wind wave, like the diurnal wave, follows the axis of the valley closely and produces a maximal supplementary wind out of the N at 05:00 DSZ. This weakens the down-valley wind flow, which is dominant at this time. Conversely, the up-valley wind experiences its maximum intensification at 17:00 DSZ as a result of the semi-diurnal supplemental term.

The rather late onset of the up-valley wind, analogous to the onset of the down-valley wind, can now be explained as the effect of the semi-diurnal term, which is distorted along the valley axis and inhibits the full development of the up-valley wind until after 14:00. Under the influence of the diurnal term only, the onset would have come before noon DSZ. The addition of the 12-hour term delayed the onset by more than one hour to 13:00.

The phase of the diurnal term of the northern component in the upper layer (1700 - 2500 m) is displaced exactly 90° compared to the phase of the lower layer. In the lower layer the extreme values of the northern component occur at 06:00 and 18:00. The extreme values in the upper layer are delayed by 6 hours. The diurnal windshifts in the upper layer are oriented essentially the same as below. The major axis of the elliptic loop, which is about four times as big as the minor axis, lies in the direction of the valley axis. Later we will discuss the consequences of this coupling of the diurnal waves of the upper and lower winds.

Nothing definite can be stated about the significance of the small second term of the diurnal cycle in the upper layer, which represents chiefly an oscillation in a WSW-ENE direction. Presumably it is of a formal nature only. In order to obtain a better approximation of the actual wind figure (compare a and d in Figure 10), a series of additional terms of the harmonic analysis would be necessary. The real meaning of these terms would, however, be highly questionable.

Daily Paths of the Wind

The above discussion considered the diurnal air transport in two distinct layers of basically different nature. To gain an even deeper insight into the diurnal wind cycle in the free atmosphere above the valley, we will now look at the relationships separately for several height levels. Specifically, we will look at the diurnal paths of parcels of air. These are less complicated than vector figures. Figure 11 shows paths for heights from 200 to 2200 m a.g.l., after removal of the daily average.

The air movement at the first two height levels, 200 m and 600 m a.g.l. is

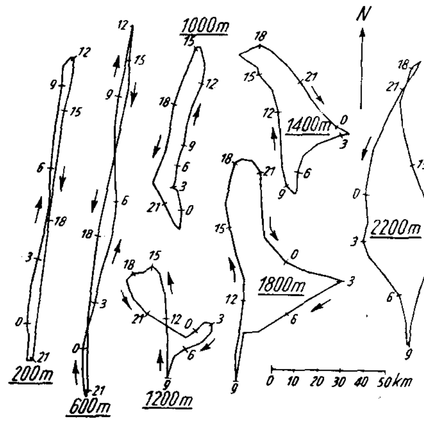

Figure 11. Movement of air mass above Sulzau illustrated by the pure periodic term over a 24 hour period. Unit in km.

essentially the same, except for differences in magnitude. There is a single periodic backward and forward shift of the air parcels along the valley axis during the course of 24 hours. The reversal points are at 13:00 and 22:00. The 600 m height represents the approximate location of maximum elongation. At the next height, 1000 m, the extent of the air movement is already considerably reduced, which corresponds to the decrease in wind speed with height. In addition, the times of the wind reversals are somewhat delayed as compared with the times for the lower layers. The reversal points of the daily path are one hour later (14:00 and 23:00) than below. The daily air movement still takes place almost exactly along the valley axis. This changes from 1200 m upwards only in so far as cross winds exert a stronger influence, especially at night. However, this does not affect the predominant N-S orientation of the air paths very much. The delay of the wind reversal continues to increase. The air at this height doesn't begin to flow into the valley until 15:00 and doesn't reverse to a down-valley flow until 09:00. Between 23:00 and 02:00, the path is oriented essentially across the valley and temporarily shows a slight down-valley component. This is apparently an intermediate layer in which the transition from the lower undisturbed valley wind cycle to an upper system takes place. This upper system continues to develop at 1400, 1800 and 2200 m. As expected, this system, too, is characterized by a diurnal periodic air displacement primarily along the valley axis. However, the wind reversal occurred at completely different times than in the lower valley regime. The wind blows towards the mountains from about 17:00 or 18:00 to about 08:00 or 09:00 and away from the mountains the rest of the time. The influence of the lower valley wind system appears to extend up to 1800 m, the height of the highest mountain peak on either side of the valley, if the noctural bulges of the 1400 m and 1800 m paths can be associated with the brief interval of the down-valley wind at midnight at 1200 m. The diurnal cycle of the upper wind system is seen most clearly at the 2200 m level.

There is no doubt about the nature of this upper wind system. It cannot be the "general" system, caused by the daily pressure wave cycling around the earth, because the curves of the paths are much too elongated in the mountain - plain direction. In our case, this direction coincides with the course of the valley, since the Salzach is a cross valley of the Alps. However, this does not preclude a secondary influence of the "general" system.

The decisive contribution to the daily cycle of the gradient wind above crest altitude at Sulzau is probably made by the compensation flow, which represents the upper link of the mountain-plains circulation. The existence of this compensation flow was proven beyond a doubt for the entire Alpine region by the research of Burger and Ekhart [10]. Up to now this proof was possible only by statistical means, using observational material extended over a rather long period of time. Because the flow is generally weak, its existence could not be proven for individual cases. The compensation flows are weak because the mass transported by weak winds in the large area above crest altitude balances the mass transported by stronger winds in the smaller cross-sectional area of the valley below.

Only in the case of cross valleys does the upper part of the circulation show considerable intensity. In cross valleys, the upward and downward flows move in opposite directions, one above the other, within the cross section of the valley. Thus, in the study mentioned above for Trient in South Tyrol by Burger and Ekhart, a return flow at upper levels was calculated which exceeded in intensity the derived mean value of the compensation flow for the entire Alpine region by a factor of 10. Wagner [2, p. 419] explains that the return flow "must occur to a large extent within the realm of the Etsch Valley itself."

Therefore it seems reasonable that in our cross valley, too, the upper compensation flow was well enough developed to be observed by a series of ascents made in not quite two days under especially favorable weather conditions. Further, this individual case offers the possibility of determining characteristic development phases of the valley wind circulation as a whole.

The Six Phases of the Valley Wind Circulation

Here again, it is sufficient to consider only the northern component of the wind, which is the controlling factor. Further, we can again refer to the two layers, which we have recognized as representative of the upper and lower branches of the circulation. Thus, the scheme gains clarity without losing essential characteristics of the phenomenon.

The upper part of Figure 12 contains the mean hourly values of the northern component of the wind connected by curves as deviations from the daily mean for the 0-1000-m layer (solid line) and for the 1700-2500-m layer (dashed line).

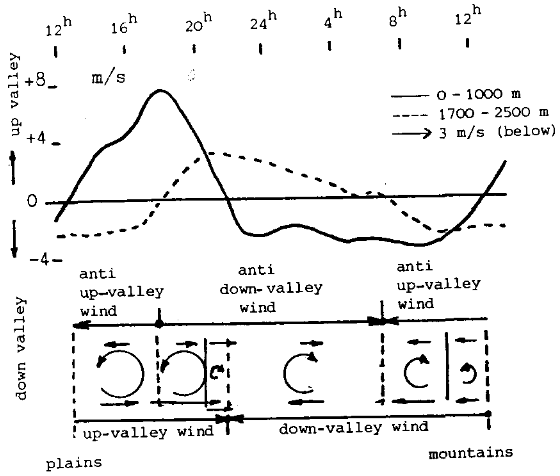

Figure 12. Diagram of the life cycle of valley wind circulation. Above - Diurnal cycle of the valley component of the wind in two layers above and below crest altitude (as deviation from the daily mean). Below: The six characteristic stages of the genesis of the valley wind, derived from the diurnal cycle of the valley component of the wind. The uppermost and lowermost arrows indicate the direction of the anti valley wind and the valley wind. The inner straight arrows indicate by their length (scale at the upper right) the mean intensity of the two flows in the time intervals indicated by the vertical lines. The circular arrows (in the middle) show in relative units the circulation resulting from the differences in the flow above and below.

We remember that positive values of the northern component indicate a flow directed from the plains toward the mountains ("up-valley") in a purely periodic diurnal cycle. In contrast, the S-component ("down-valley") indicates a wind blowing from the mountains toward the plains. One notices immediately that the upper wind and the lower wind, as we will name the flows in the relevant layers for now, blow towards each other during the greater part of the day. For approximately 9 hours (22:00 - 07:00) a N-wind prevails above, a S-wind below. Conversely, for 5 hours (13:00 - 18:00) a N-wind prevails below and a S-wind above. There are then 10 hours during which the two flows have the same direction. This is surprising at first. However, if one considers the development phases through which a thermal circulation passes to a stationary state, according to the scheme developed by Hann, one arives with Moeller [11, p. 506ff] at the following picture: "Since the movement of the air at upper levels takes place almost without friction, but the movement of air at lower levels near the earth's surface is greatly retarded by friction, the air at upper levels usually precedes the air at lower levels by quite a bit. The air pressure falls (or rises) at the earth's surface without the initiation of the corresponding air movements. A greater pressure difference must first develop, before it can overcome the obstacles to motion." The retardation of air movement caused by friction is no doubt even greater for flows in mountain valleys because of the addition of friction along the sidewalls.

To what extent the motion at upper levels precedes the motion it initiates below can be read directly from Figure 12. I believe this applies not only to this case, but also for the first time to this type of thermal circulation in general. After 18:00 the wind at upper levels is already blowing from the plains toward the mountains. However, not until 22:00 is the mountain-plain pressure gradient great enough within the valley for the down-valley wind to overcome the retarding force of friction. Therefore the "anti down-valley wind", as the part of the circulation at upper levels should obviously be called, begins four hours earlier than the down-valley wind. It is the same for the up-valley wind, which follows the "anti up-valley wind" with an even greater delay of 6 hours. The domains of the anti valley wind and the valley wind therefore overlap in time, so that there is already a natural division of the development of the valley wind circulation into four phases. If we add the two times at which the upper wind and the lower wind of the circulation are the same strength (intersections of the two curves, Figure 12, above), a division of the valley wind development into a total of 6 characteristic stages results. These six stages are illustrated in the lower part of Figure 12, which, in connection with the diurnal cycle of the valley wind components shown in the upper part of the figure, is self explanatory. The figure can be discussed as follows.

Regime of the Up-valley Wind (13:00 - 22:00)

Phase 1 (13:00 - 18:00): The up-valley wind is beginning below. At upper levels, the anti up-valley wind (compare phase 6) is already abating. The result is a circulation as indicated by the circular arrow[a], with ascending motion in the warm mountains and descending motion over the cooler plains.

Phase 2 (18:00 - 21:00); An anti down-valley wind begins at higher levels at the same time the up-valley wind below is fully developed. The counterclockwise circulation continues with the same intensity as the result of an almost equal relative velocity difference between above and below. (The circular arrow is almost exactly the same size as in Phase 1).

[a] In the figure, the size of the circular arrows indicates the varying intensity of the circulation in the individual phases resulting from the changing difference in velocity between above and below.

Phase 3 (21:00 - 22:00): While the anti down-valley wind increases in strength at higher levels, the up-valley wind below weakens noticeably. This results in a circulation tendency, the reverse of that in Phases 1 and 2, with descending motion in the mountains and ascending motion over the now warmer plains. Thus, this circulation tendency works against the prevailing circulation of the plains-mountains system. The up-valley wind below must then cease, resulting in the development of ....

The Regime of the Down-Valley Wind (22:00 - 13:00)

Phase 4 (22:00 - 07:00): There is a down-valley wind below and an anti down-valley wind above. The wind speed above is lower than below. Therefore the circulation introduced in the previous phase is strengthened.

Phase 5 (07:00 - 10:30): The down-valley wind is somewhat stronger below. In contrast, the anti up-valley wind is already beginning at higher levels. Since the anti up-valley wind is still weaker than the down-valley wind, the circulation continues, but is already noticeably weaker than before, because both the upper and lower winds are blowing in the same direction.

Phase 6 (10:30 - 13:00): A strengthening of the anti up-valley wind and a simultaneous weakening of the down-valley wind to below the absolute velocity value of the anti up-valley wind results in another reversal of the direction of circulation. It is now warmer in the mountains than over the plains again, resulting in a descending tendency in the mountains and an ascending tendency over the plains. This is the "dying phase" of the down-valley wind, and the process now begins again from the beginning (Phase 1).

The Compensation Flow

The question now arises, whether it is possible to go beyond this general scheme to make quantitative statements about the extent and intensity of the circulations in the quasi-stationary states. This is difficult since the local wind system does not exist in isolation. Primarily above the crest altitude, it is overlain by flows which result from the large scale pressure distribution and which are not constant with height. However, because we only need to consider one component of the wind, the valley component, an attempt can be made to answer the above question.

Figure 13, left diagram, shows the velocity profile of the N-component under down-valley wind conditions over Sulzau (thick line). This was used earlier, but is extended here to higher levels. The lower shaded surface between this curve and the ordinate axis shows the regime of a negative N-component, or of the down-valley wind, if the influence of the gradient wind at these levels can be precluded, as we determined earlier. Above this the

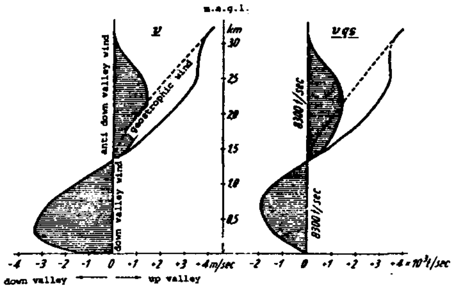

Figure 13. Determination of thickness and strength of the compensatory flow of the down-valley wind (Sulzau). Left: height distribution of the valley component of the wind and division of the upper air wind into geostrophic wind and anti down-valley wind (m/s). Right: mass flux of the flow in the direction of the valley. The two hatched areas are of equal size (down-valley wind below and anti down-valley wind above) (10^3 tons/s).

N-component has a positive sign. On the one hand, this is the gradient wind, on the other it is the return flow of the lower wind, which has to have an upper boundary. The effective range of the N-component of the geostrophic wind should begin just at the upper boundary of the down-valley wind according to our conditions mentioned above. In our case, this would be a little above 1300 m. From here on up, its strength should increase steadily.

The diminished vertical increase of the velocity, clearly discernible in the figure between 2500 m and 3000 m, seems to me important. Although not provable, it seems to be obvious and plausible to relate this to the upper boundary of the compensation flow, similar to the case in Innsbruck [12], and to delineate the compensation flow as the local part of the wind, as is shown in the figure. This would result in a layer thickness of about 1800 m (1330 - 3160 m a.g.l.) for the anti down-valley wind, i.e., almost 50% more than for the down-valley wind.

Of course, it is not known at present in what way the velocity of the compensation flow changes within its height range, since we have only the sum of the gradient flow and the compensation flow in our diagram. However, we know that within the regime of the valley wind circulation the total mass flux through a vertical plane perpendicular to the direction of flow, i.e., in our case perpendicular to the valley, must be zero. In other words, the air in the valley below which is transported out of the mountains by the down-valley wind must be replaced above by the return flow from the plains.

This condition was used in the right half of Figure 13 in the calculation of the product of velocity v times the flow cross section q times thickness, resulting in the division for the velocity distribution shown by the two curves (dashed and thinly drawn) in the left diagram. It must still be noted that the upper valley width was used as cross section q for the flow above the ridge top level, which is no doubt a simplification, but probably a permissible one. In this way the height and velocity distribution of the compensation flow, which is also shown by shading in the left part of Figure 13, was determined. Its maximum intensity is at about 2250 m a.g.l. and compares to that of the down-valley wind below in the ratio of about 1:2 (1.5 to 3.3 m/s), which appears to be quite reasonable.

It is not possible to make a similar calculation for the up-valley wind, since the relationships are no longer so clear and simple due to the orientation of the gradient wind and the unstable thermal stratification.

REFERENCES

1. Wagner, A., Neue Theorie des Berg- und Talwindes [New Theory of the Mountain and Valley Wind]. Met. Z. (1932) 329.

2. Wagner, A., Theorie und Beobachtung der periodischen Gebirgswinde [Theory and Observation of Periodic Mountain Winds]. Gerl. Beitr. 52 (1938) 408.

3. Ekhart, E., Zur Aerologie des Berg- und Talwindes [On the Aerology of the Mountain and Valley Wind]. Beitr. Phys. fr. Atm. 18 (1932) 1.

4. Ekhart, E., Die Windverhaeltnisse des Arlbergs II. Teil: Aerologische Beobachtungen [Winds of the Arlberg; Part II: Aerologic Observations]. Gerl. Beitr. 49 (1936) 7.

5. Ekhart, E., Ueber einige Gletscherwindmessungen in den Oetztaler Alpen [About Some Measurements of Glacier Winds in the Oetztal Alps]. Z. Gletscherkde. 22 (1935) 217.

   ________. Neuere Untersuchungen zur Aerologie der Talwinde: Die periodischen Tageswinde in einem Quertal der Alpen [New Investigations of the Aerology of the Valley Wind: the Periodic Daily Winds in a Cross Valley of the Alps]. Beitr. Phys. fr. Atm. 21 (1934) 245.

6. Moll, E., Ueber die Windverhaeltnisse des Engadins, speziell den Malojawind [The Winds of the Engadin, Especially the Maloja Wind]. Beitr. Phys. fr. Atm. 24 (1938) 270.

7. Jelinek, A., Untersuchung periodischer Tageswinde in Suedtirol [Investigations of Periodic Daily Winds in South Tyrol]. Beitr. Phys. fr. Atm. 21 (1934) 223.

8. Pollack, O., Der Foehn in Salzburg [The Foehn in Salzburg]. Progr. k.k. Straatsgymn. Salzburg. 1909/1910.

9. Schmauss, A., Foehn in Partenkirchen [Foehn in Partenkirchen]. D. Met. Jb. Bayern (1926) Appendix E.

10. Burger, A. and E. Ekhart, Ueber die taegliche Zirkulation der Atmosphaere im Bereiche der Alpen [The Daily Atmospheric Circulation in the Alps]. Gerl. Beitr. 49 (1937) 341.

11. Hann, Meteorologie, [Meteorology, 5th edition].

12. Ekhart, E., Die Windverhaeltnisse Ueber einer inneralpinen Station [The Wind Situation at an Inner Alpine Station]. Beitr. Phys. fr. Atm. 23 (1936) 165.

ON THE THERMAL STRUCTURE OF THE MOUNTAIN ATMOSPHERE

E. Ekhart

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ON THE THERMAL STRUCTURE OF THE MOUNTAIN ATMOSPHERE

E. Ekhart

Temperature measurements made horizontally and vertically in a valley and above a plain permit for the first time the direct testing of hypotheses set forth on the subject of the origin of the diurnal air circulation between the mountains and the plains. Differences in the thermal structure of mountain and valley winds are brought into evidence. The evidence of the double structure of the diurnal thermal wave above the valley is given, which is a new result. The analysis of the diurnal variation of the temperature in the free mountain atmosphere and the facts already known, such as the specific phenomena of mountain circulations, leads to a diagram of the discontinuous structure of the atmosphere in the region of a mountain.

Diurnal Variation of the Temperature at the Ground

Following the new theory of Wagner the up-valley wind represents a local invasion of cold air into the valleys and thus causes the lowering of the mean maximum temperatures, in particular of the lowest part of the valley, when the cold air coming from the plains hasn't yet had time to warm sufficiently [1, p. 332].[1] And a little later one reads:

"Thus the mountain and valley wind system is entirely analogous to the land-sea breeze system."

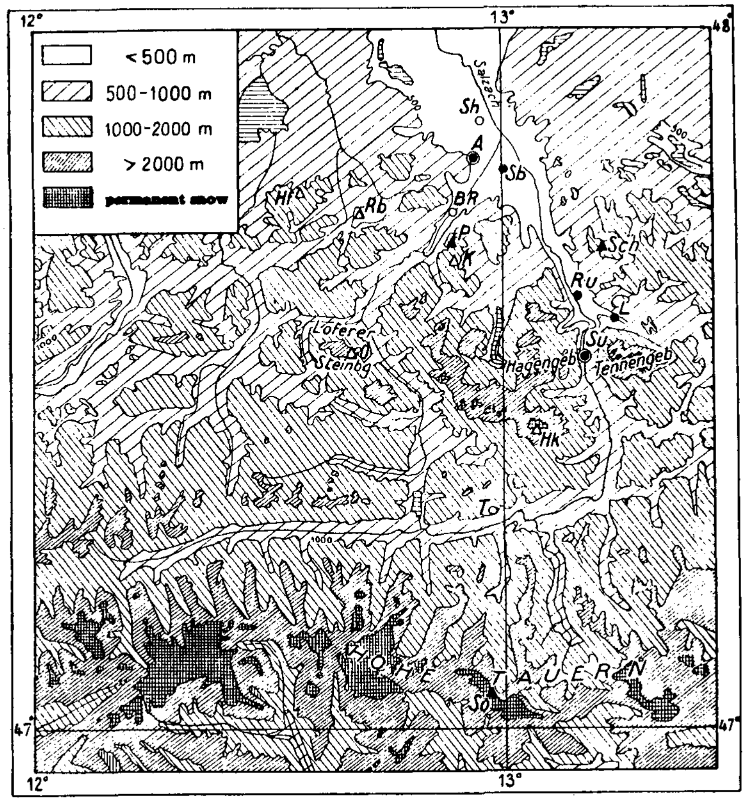
As for the land-sea breeze system, we currently know the characteristic recordings of temperature and humidity produced at the coastal stations, in particular the rapid fall of the temperature, more or less accentuated according to conditions, and the corresponding rise of relative humidity which indicates the arrival of cold marine air. To the extent that one travels towards the interior, this thermal symptom becomes less perceptible by virtue of heating and mixing of the marine current with continental air. Wagner assumed that something analogous takes place for the valley wind.

Indeed, while looking at the thermogram of a valley station in the interior of the Alps, where the valley wind is clearly characterized, one has the firm impression that the diurnal maximum seems to be "lowered". Nevertheless, this fact is not directly accounted for by the invasion of colder air coming from the plain. In fact the more intense mixing of the valley air due to the augmentation of the force of the wind--and consequently of the dynamic exchange--when the up-valley wind begins to blow, could explain this fact without need for the supplementary hypothesis of advective phenomena.

But if one considers thermograms for beautiful summer days at a mountain station situated at the entrance to a valley one finds Wagner's point of view completely confirmed. In support of this thesis we present in Fig. 2 copies of temperature and humidity observations taken at Salzburg[2] during the period of observations considered, characterized by a particularly typical time of valley wind. This fact is manifest in the extremely uniform daily

[1] The numbers in brackets refer to the references given at the end of the article.

[2] For geographical information and the localities cited in the text, see Fig. 1. Time is indicated in German Summer Time (D.S.Z.) - Central Europe Time plus 1 hour.



AEROLOGIC STATIONS

A _ *Ainring* 457 m. (radio soundings)
_ *Munich* 530 m. (aircraft; not shown on the map)
Su _ *Sulzau* 510 m (radio soundings)

THERMOMETRIC OBSREVATION POSTS

L _ *Lehngries* 510 m.
P _ *Predigtstuhl* 1576 m.
Ru _ *Russeck* 470 m.
Sb _ *Salsbourg*-Airport 434 m.
Sch _ *Schlenken* 1649 m.
So _ *Sonnblick* 3106 m.

OTHER LOCALITIES CITED IN THE TEXT

BR _ *Bain Reichenhall* 465 m.
Hb _ *Hochfelln* 1658 m.
Hk _ *Hochkönig* 2938 m.
K _ *Karkopf* 1739 m.
O _ *Grosses Ochsenhorn* 1543 m.
Rb _ *Rauschberg* 1646 m.
Sh _ *Surheim* 430 m.
T _ *Taxenbach* 1100 m.

Fig. 1. Map of the region studied: Salzach Valley (Eastern Alps)

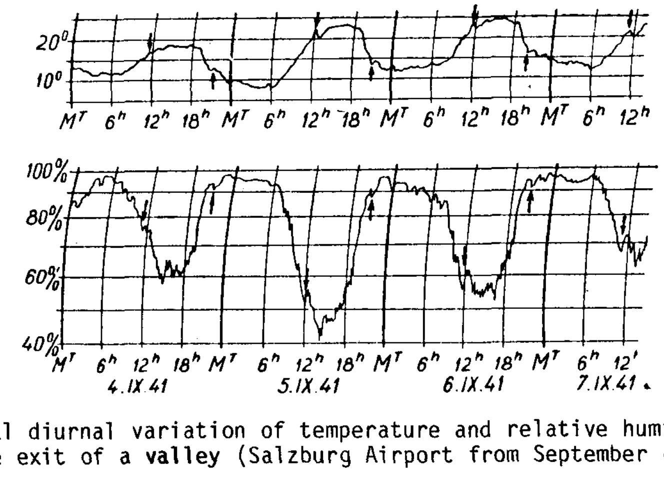


Fig. 2. Typical diurnal variation of temperature and relative humidity of valley wind at the exit of a **valley** (Salzburg Airport from September 4 to 7, 1941)

The arrows indicate the start of the up-valley wind (↑) and of the down-valley wind (↓) .

variation of consecutive days with strong periodic variations, characteristic for a station situated in a valley.

The curves which have a perfectly regular appearance show again a particular characteristic in consequence of a phenomenon which--as far as I am aware--isn't known for any of our Alpine stations. The relatively rapid rise of temperature in the forenoon is suddenly stopped near noon by a fall in temperature (↓) of about 2°C, followed shortly by a nearly equivalent rise. But after the break in the diurnal curve, the curve does not resume its earlier rate of rise. Instead, one gets the impression that starting from the moment of the fall of temperature the crest of the curve is broken and slips to a lower level. One doesn't need much imagination in order to be able to picture how the variation of the curve would be formed without this perturbation.

This drop of temperature, changing the appearance of the curve and, during good weather appearing regularly at noon, is a manifestation of the invasion of colder air coming from the plain and advancing towards the mountains. This air flows in the form of an up-valley wind of the Salzach valley. We don't have at our command the anemometer records needed to verify this phenomenon. However, from what we know of the up-valley wind in the Salzach valley, notably the hour at which it occurs (compare [2]), and also taking account of personal observational reports, there is no doubt of the accuracy of this statement. At this point, situated at the entrance to the mountains, the coldest air of the plains still exhibits much of the specific properties of its place of origin. These properties are modified because of the processes of mixing and radiation as the air flows up the Salzach valley; this is placed in evidence by the curves of Fig. 3, deduced from readings taken with an Assmann psychrometer every 15 to 30 minutes at the Russeck observation post, situated 20 kilometers from the entrance to the valley. On the other hand, the up-valley wind cannot be considered as a more-or-less isolated particle of air in conformance with earlier studies (see, for example, [2, p. 225]). Moreover, the fact that the up-valley wind appears practically at the same time along the whole length of the valley is already a proof to the contrary.

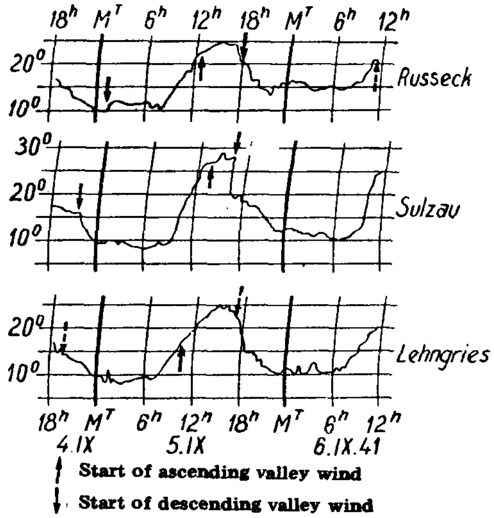


Fig. 3. Diurnal temperature variation at the observation posts (according to readings with the Assmann psychrometer at 15-30 minute intervals).

If one studies the Salzburg thermograph more carefully in order to study what occurs at the moment when the evening wind changes at the time of the arrival of the down-valley wind, which introduces into the system a mass of cold air directed from the mountains towards the plains, one ascertains an inverse effect from that at noon (the instant that the up-valley wind begins). This effect is less noticeable, but is sufficiently characteristic, i.e. heating (↑). When the down-valley wind enters into the plain it has already undergone a long journey. The effect of a cooling will then be limited to simple local radiation. Indeed Wagner (*loc. cit.*) supports the view that the cold mountain wind makes the minimum temperature lower but he adds that this occurs "above all, far into the vast plain into which the valley opens". Besides, without the knowledge of particular atmospheric phenomena, we would have no obvious reason to assume the contrary effect, all the more so because the heating not only doesn't diminish in the first hours of the night, but effectively causes a rise of nocturnal temperature (see in particular the night of 5 to 6 September). The explanation is given by the spreading of the cold air at the exit of the valley into the vast plain, a phenomenon normally associated with dynamic heating. This explanation has already been given [2, p. 225].

It is the particularly intense displacement of air at the exit of the valley by the down-valley wind, preventing the cold air at the ground from being stagnant, which results in an agitated nocturnal temperature variation. This is in general the case at the interior of large and flat valleys such as the Inn and the Drave valleys.

It is unnecessary to stress the fact that the peculiarities of the thermogram are reflected in the records of relative humidity.

The more we advance into the valley, the more the increase in the diurnal temperature variation (Fig. 3). If we choose as an example the rise of temperature from the minimum of the night of the 4th and 5th of September to the maximum of the 5th, we find at Salzburg 15.3°C, at Russeck 14.9°C, at Lehngries 17.2°C, and at Sulzau 20.8°C. The results established for what occurs at the ground remain exact for the entire atmospheric layer between the bottom of the valley and the height of the neighboring crests. It is on this fact that the new valley wind theory of Wagner ultimately rests.

Apart from this and except for the already observed short variations during the night, nothing appears remarkable in the temperature curves of Russeck and Lehngries. In particular, they no longer exhibit traces of "frontal symptoms" at the moment when the wind begins to turn, as was the case for Salzburg[1].

In contrast, the Sulzau thermogram provides a typical example of conditions at a constriction of a valley. Not only (as we have already indicated) is the diurnal variation extremely large in this region, but the air enclosed between the sloping valley sides reacts much more rapidly to radiation phenomenon than elsewhere. In consequence of the later rising of the sun the morning heating begins a little later than at the other valley stations, but this delay is quickly recovered thanks to the extremely rapid heating in the morning. The smaller the mass of air to be heated, the more the temperature variation follows that of insolation. It is for this reason that, in the afternoon, almost immediately after sunset, an appreciable drop of temperature occurs, which leads to the formation of an inversion at the ground. However, as we will mention later, this drop is limited to a thin layer near the ground (on the 5th of September, a 10°C drop occurred in a 1/2-hour interval and a 7°C drop occurred in a quarter hour). Such phenomena approach the limits of macro- and microclimatology.

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[1] For the reason of simple curiosity we draw attention in particular to the Lehngries thermogram on the point of rise after midnight. This presents a regular anomaly of the variation of nocturnal temperature. Such "anomalies" are well-known by the detailed study of Huber [3].

Figure 4 will serve as an example of the difference of daily temperature variation at the ground due to site location. The average daily variation of the temperature difference between Sulzau and Salzburg as deduced from observations taken on the 4th through the 6th of September 1941 is represented in this figure. Note that Lamont's correction has been applied. In comparison to Salzburg the daily average at

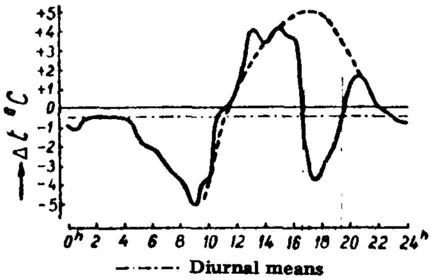

Fig. 4. Variation of the difference between the temperature of two stations in the interior of the valley and at the exit of the valley (Sulzau-Salzburg)

Sulzau is a little lower, or in more precise terms: during the up-valley wind the temperature is visibly more elevated--4°C at the maximum--and at night at the time of down-valley winds, much lower--5°C at the maximum. We would obtain the same figure, at least qualitatively, if we had observations of the whole column of air above the valley. It is this that we will prove in the following sections. It is only the transitory drop between 1645 and 1939 hours which would disappear in the difference curve, because it is caused by the abrupt temperature fall at the end of the afternoon only in the neighborhood of the ground at Sulzau. The probable temperature curve, setting aside this perturbation, has been marked by dashed lines on the curve.

Daily Variation of Temperature in the Free Atmosphere in Mountains

All of the studies made up to now on the daily variation of temperature in the Alps were based on the observations of fixed mountain stations. But, even in situations with well separated summits, the influence of proximity to the ground makes itself clearly felt. The series of Sulzau radio soundings made on the 4th through the 6th of September 1941 give us an idea of the diurnal variation of temperature in the free atmosphere in the interior of a mountain range. The notion of "free atmosphere" will be defined in the last chapter.

Firstly, we show the division of atmosphere according to the hour of the day and the altitude above sea level based on 13 soundings in the form of carefully analyzed actual and potential temperature isopleths (Fig. 5). The time of the soundings has been indicated by arrows at the base of the figure.

On the whole, thanks to the undisturbed atmospheric situation during the period considered, the figure gives the impression of a rather balanced set of curves, bringing out the diurnal temperature variation. It is only at the beginning of the series of observations, the afternoon of the 4th, when greater perturbations occurred. These were certainly due to the general meteorological situation, since they are also found in the wind isopleths of [2, Fig. 2].

The surprising thing is the very marked daily variation in the whole valley cross-section, characterized by the dome-shaped isotherms n the afternoon of the 5th. In the region near the ground the maximum temperature is attained at 1500 hours. The isopleths of potential temperature show that a dry-adiabatic gradient then prevails at the interior of an atmospheric column of nearly 2000 meters. Immediately after, rapid stabilization of the valley air is established, progressing

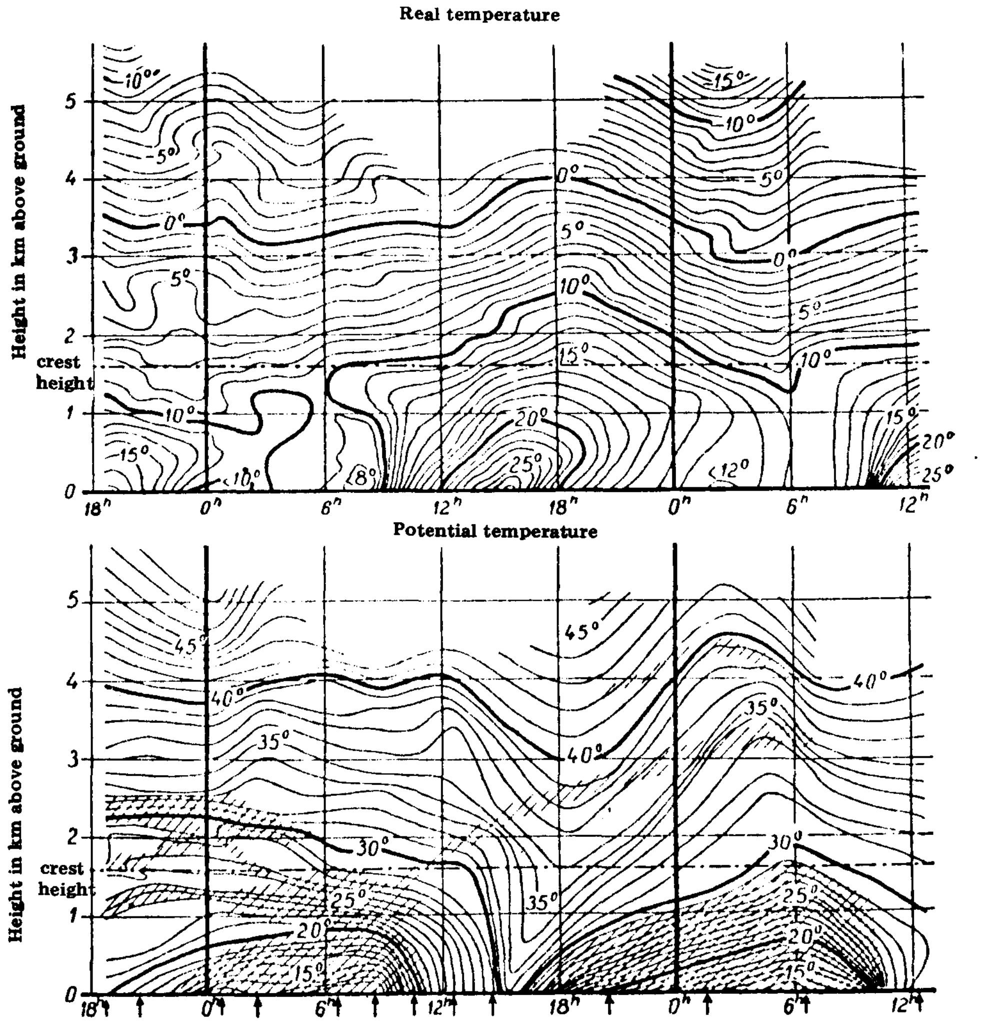

Figure 5. Isotherms of the real temperature (top) and the potential temperature (bottom), according to radio soundings in an Alpine valley (Sulzau-s/Salzach). Hours of soundings indicated by arrows at the bottom edge of the figure. The hatched zones of the bottom figure designate layers with inversion or very weak temperature gradients.

from the bottom to the top of the column. The lowest layers cool rapidly while the temperature above is still rising slightly. At the altitude of the ridge tops (1600 meters above the ground or 2100 meters above sea level) the daily maximum occurs between 1800 and 1900 hours. A very strong inversion is produced (hatched area at the bottom of Fig. 5). At first, this occurs at the ground in the form of a thin layer, then constantly gains in depth and attains the altitude of 1200 meters a little after midnight. This depth is only 400 meters from the edges of the lateral plateaus. In an analogous fashion, but delayed by a stronger atmospheric disturbance, an inversion occurred above the valley during the previous night (4-5 September) and attained a height of 1300 meters above the ground at the end of the morning of the 5th. In both cases it was rapidly dissipated. On the 6th, this occurred nearly all-at-once in the first 1000 meters, the atmospheric layer having become unstable following the heating of the ground. In the course of the 2 mornings the mean temperature of the valley atmospheric column (0 to 1600 m) rose in four hours (8 to 12 am) by 7°C, and on the 6th at noon the structure of the first 500 meters became unstable in comparison with dry air.

Not only is the diurnal variation in the whole valley cross section incomparably stronger than in the layers of equal thickness over the plain, but the appearance of the isotherms shows clearly that the diurnal temperature variation extends well above the crests. It has been observed even to a height of 6000 meters above sea level, the highest level attained in the soundings. However, associated with this variation is a surprising fact concerning the variation of phase with height. Consider for example the hour of the daily maximum at different altitudes. On the 5th we first observed the already mentioned delay of the phase as a function of height (ground: 1530; 1600 m: 1800 hours). This delay is less accentuated above the crests and disappears nearly completely above 2500 m. The morning minimum has a rather inverse tendency, namely the advance of the phase with elevation.

In order to bring out the periodic variation, the variation of temperature of the 5th of September--a day nearly without perturbations--has been specially studied and represented in the form of isopleths (Fig. 6) after elimination of aperiodic variations. The upper part of Fig. 6 represents absolute values. The lower part represents deviations from the daily mean. The analysis of data was limited to 4000 m above the ground in order to minimize any potential problems due to interpolation effects. There were sufficient observations below 4000 meters to allow a good analysis.

Fig. 6 clearly shows the most characteristic traits of the whole phenomenon. The result is rather unexpected: <u>A splitting of the vertical structure of the diurnal temperature wave in the mountain atmosphere</u>. We find the usual form of the diurnal period from the ground to the altitude of the ridge tops, in which the amplitude decreases and the phase is delayed with height. This feature is clearly characterized in this particular example. Above the ridge top heights the deviations from the daily average are weakest and at this altitude the diurnal variation is least developed (amplitude only about 3°C). But still higher above, between 2-1/2 and 3 km above ground, the amplitude increases again and attains a secondary maximum of nearly 6°C [*sic*, 4°C]. Thus, <u>the atmosphere above the level of the mountain crests has its own diurnal variation, which is very different from that of the atmospheric layer enclosed in the valley and situated below</u>. This new result is extremely interesting.

Since this concerns the dominant first term of Fourier's series the relationship can be best tested in the form of the hourly periodogram represented in Fig. 7. In the atmospheric column bounded by the ground and the height of the crests, one can distinguish two layers. In the lower layer, about 400 m above the ground of the valley, the amplitude is practically constant (about 7°C); in contrast, the phase changes greatly (by more than 1 hour). From 400 m to 1600 m the phase changes little but the amplitude decreases steadily. At 1600 m it isn't more than 2°C. This type of change of diurnal temperature variation with height at the interior of a valley can be a purely local particularity, uniquely valid for the Sulzau canyon. But what is more significant is the second part of the vector curve for the diurnal wave. After an abrupt phase delay, the amplitude increases anew after 1700 m and attains a second maximum of about 3°C at 2800 m. It is only above this altitude that the

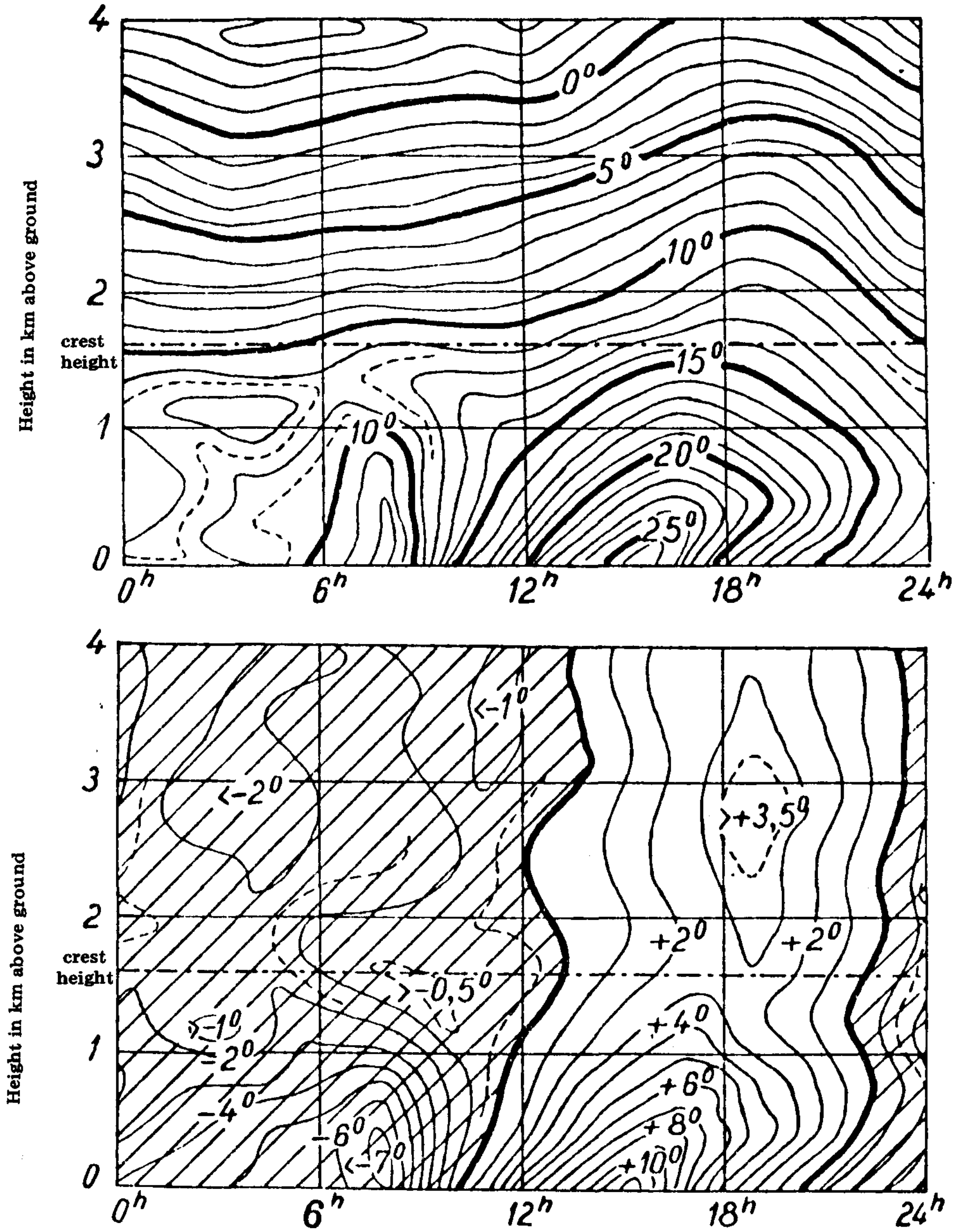

Figure 6: Diurnal temperature variation in the free atmosphere above an Alpine valley station (Sulzau, September 5, 1941).

Top: isotherms of the real temperature corrected for the differences between 2400 and 0000 hours.
Bottom: differences from the diurnal mean.
Hatchings: below the mean.

amplitude continues to decrease. Between 3000 and 4000 m the values are brought into agreement. The amplitude then slightly exceeds 2°C (the radiation term of W. Schmidt, [4]), and the maximum occurs at about 1800 hours.

Indeed it would be rather hazardous to seek to make general conclusions about the mountain atmosphere temperature variation by basing them on a single series of observations made at a single station. Yet it would be just as arbitrary to consider an interesting result as due to pure chance. For the case of daily variation of meteorological phenomena, we are well aware that it is often possible to identify what is

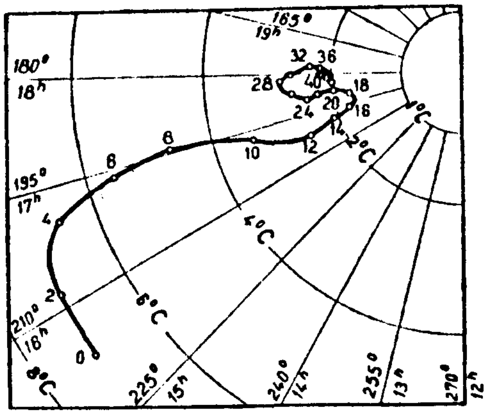

Figure 7: Hourly diagram of the periodicity of the diurnal thermal wave above a valley station (Sulzau). Numbers given at different points of the curve are heights in hectometers above the ground.

typical by limiting ourselves to a few cases, on the condition that they are sufficiently representative for the phenomenon in question. The day chosen fulfills this condition. Moreover, the physical reality of a phenomenon always becomes more probable to the extent that we can give a more plausible explanation for it. On the other hand the most powerful mathematical formalism has never had the power to prove the physical existence of a phenomenon (example: correlation).

In the case at hand this natural interpretation seems to me to be valid. The double structure which occurs so clearly in the daily temperature variation evidently indicates that two quite different processes are to be considered in its formation. The process of diurnal heating of the air in the valley itself doesn't require a particular explanation. Its intensity and its very great extent in altitude in comparison to conditions above the plain issue from the increase in area of the heated surface (valley floor and slopes) and from its "effective" altitude which is higher in comparison to conditions above the plain, even in the case where the valley floor and the plain are at equal levels. By virtue of heating and radiation the same volume of air receives more heat during the day and loses more at night in the valley than over the plains. Furthermore, it must be assumed (following Wagner) that convective mixing is noticeably reinforced in the valley relative to the plain. Following Wagner, the upper limit of this perturbed zone that is slightly warmed during the day slightly surpasses the ridgecrests because of turbulent exchange and radiation.

At a certain height above the ridgecrests the thermal influence the valley air partially disappears. By contrast the mountain chain acts, during the day, as an elevated heat source, detached from its flat surroundings [5, p. 62], so that a particular daily temperature variation can be established above it independent of what occurs below in the troughs (valleys and smaller valleys) of the mountainous region. It is really this that is shown in the observations by the secondary maximum of the daily temperature wave near 2800 m above the ground or 3300 m above sea level.

For the moment we must wait to find out to what extent these results concerning the diurnal variation of temperature in the mountains can be generalized. Because of these results, our present conceptions may be susceptible to modification. However, at this moment these results can be supported by the observations made by K. Brocks [6], obtained in a very different manner. Four years ago Brocks had studied the atmospheric structure above different parts of the Northern mountain chain of the Berchtesgaden Alps from the summit of the Karkopf (1739 m) by means of refraction measurements. His studies included, among others, the study of the diurnal variation in the characteristic layers of the mountain atmosphere. The atmospheric layers which lie below 2300 m, and are thus situated below the mean ridge heights and subject to the influence of the variation of slope and valley winds, were explored by a light ray directed towards the Hochfelln (1658 m) and the Rauschberg (1646 m), which are mountains of the subsidiary mountain chain of the Chiemgau Alps. Above 2300 m the atmospheric conditions had been studied by means of sightings in the direction of the Grand Ochsenhorn (2543 m) in the Loferer Steinberger mountains, and finally at greater altitudes by sightings in the direction of the Hoch Koenig (2938 m), situated more at the interior of the Salzbourg Calcaires [Limestone] Alps. Considering the mean of measurements made during 3 entire days it seems that in the layers in the neighborhood of the crest regions the diurnal temperature amplitude increases anew, before decreasing rapidly above with altitude, presenting at the same time a delay in phase [6, p. 70]. There isn't the least trace of this second maximum in the vertical distribution of the diurnal amplitude above the plain. Brocks, in discussing the "conditions of the mountain-plain transition zone", stresses this fact, which in other respects was already known by the scrutiny of a series of soundings at Lindenberg [7].

This result, in agreement with ours, is all the more interesting since the measurements were made in other regions of the mountains and during different seasons. However, although the mountain soundings demonstrated the phenomenon, it is without doubt because it is particularly well-marked at a valley constriction confined between the Karst plateaus, which are perfect heating surfaces. The phenomenon probably doesn't occur with the same distinctness in basins or in larger valleys.

Thermal Structure in Up- and Down-Valley Winds

Aerological soundings made in valley winds permit the study of the thermal difference of the up- and down-valley winds and of the "altitude wind". When the "altitude wind" is directed across the valley wind direction, the transition from the valley wind to the geostrophic wind occurs not gradually but abruptly (compare [2, p. 221 ff]). Wagner gave an explanation of this fact by assuming the existence of a thermal separation surface at this height. This hypothesis can be easily verified by drawing a mean temperature curve for the case of an up-valley wind and for a down-valley wind by utilizing corresponding radiosonde observations. One chooses the altitude and temperature of the boundary between the valley wind and the gradient wind as the point of origin of the temperature-altitude coordinate system.

The result is interesting (Fig. 8): The air of the down-valley wind has an extremely stable structure. It is valid both for the mean and for the particular case, given that the mean slightly masks the more or less elevated or more or less developed thermal limits in the particular cases concerning the down-valley current. The system thus appears in the following fashion: In the lower layers we find a slight inversion; above, a slight decrease of temperature and in the last 100 to 150 m below the upper limit of the local wind a new inversion of temperature. On the average all of the region of the down-valley wind is ruled by near-isothermal lapse rates. But the boundary between the local wind and the "altitude wind" is characterized by a marked discontinuity of the vertical temperature gradient for, in the region of the gradient wind, the normal decrease of the temperature with altitude predominates (in our case 0.55°C/100 m).

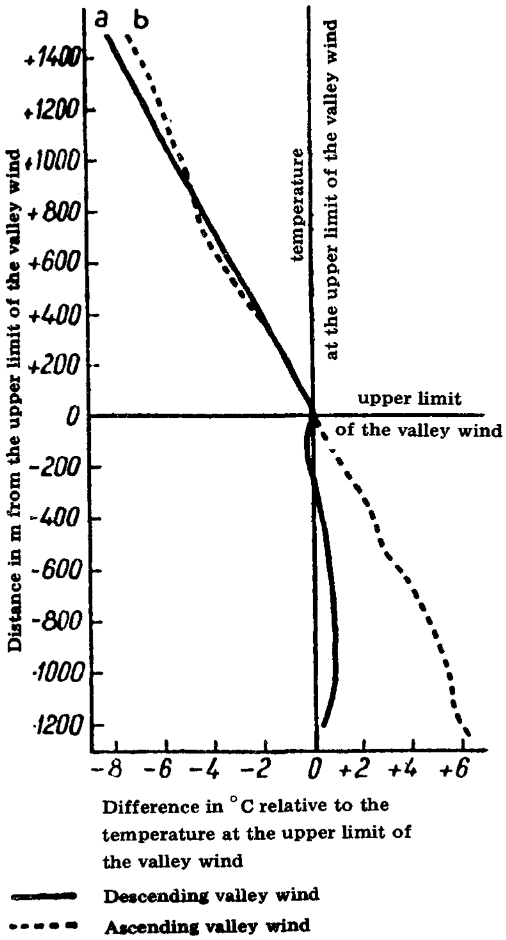

Figure 8: Mean vertical temperature distribution of down-valley wind (a) and of up-valley wind (b) above Sulzau (relative to the upper limit of the valley wind).

In contrast, the upper limit of the up-valley wind isn't at all clear in the representation of the vertical distribution of temperature, neither by a temperature discontinuity, nor by a discontinuity of temperature gradient. The temperature decreases regularly in all of the region studied and in an absolutely normal fashion. It is only at around 700 m above the wind boundary that a slight inflection of the sounding curve occurs by virture of a slackening in the decrease of temperature. Given that the thickness of the up-valley wind was 1400 m during the sounding period (that is at 1900 m above sea level), the diminution of the vertical temperature gradient at the altitude of 2600 m can only be a consequence of the heating effect by the mountain chain.

Following Exner [8], the [turbulent] exchange depends essentially on the degree of stability of the structure. Our representation thus permits us to conclude that just at the transition from the down-valley wind to the "altitude wind" the exchange coefficient increases in a discontinuous fashion. That is to say that the coupling of the lower wind to the upper wind is then relatively weak. It isn't the same, at least not to the same extent, for the up-valley wind. Perhaps in consequence of the generally greater extension in altitude of its influence (see [2, p. 219 ff]), better mixing of air masses (and thus of the amount of exchange) occurs between the two currents. The direction of this movement is determined by the mass of air in the lower layer, because its channeled flow through the valley, limited by the mountains, has less liberty of movement than the wind in the free atmosphere above the crests. All this is in perfect accord with the results already obtained by means of simultaneous soundings [2].

Thus, if we pass anew from the mean to the observations themselves, the daily course of temperature gradients below and above the ridge tops (representing approximately the upper boundary of the valley wind) must be different. Fig. 9 gives an idea of this by comparing the 400 to 1600 m layer and the 1800 to 2600 m layer. Each layer undergoes its own variation. On the whole the two curves are in opposition. Whereas, at night the valley air has weak positive gradients and even negative gradients (inversion, minimum at 0800 hours) and during the day has strongly positive gradients (maximum at 1600 hours), the atmospheric layer above the mountain has stronger gradients at night than in the

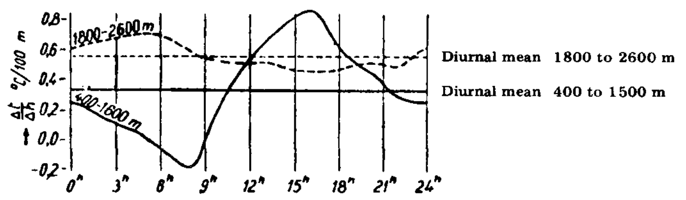

Figure 9. Diurnal variation of the vertical temperature gradient in the atmospheric column of a valley and above crest height (Sulzau).

day. At altitude the daily variation is noticeably weaker than in the lower layers. The temperature gradient of the valley air is on the average weaker than at altitude, and this is without doubt because of radiation and relatively short insolation of the N-S valley. In the valley region the diurnal variation is represented by a diurnal wave of 24 hours which still prevails clearly at altitude. But there is, in addition, the trace of an 8-hour cycle in which the maxima occur nearly at 0400, 1200 and 2000 hours, a fact in perfect agreement with the results obtained by Hergesell at Lindenberg [7]. We say this in passing.

By taking the vertical temperature gradient as an index of the intensity of the turbulent exchange, one recognizes immediately that the mixing between gradient and local winds is strongest between noon and evening, and thus during the up-valley wind regime.

<u>Temperature and Pressure Differences in the Free Atmosphere Between Mountain and Plain</u>

Thanks to the soundings made twice per day at Munich in this time period, it is possible to make a direct comparison of temperature between the free atmosphere in the mountains (Sulzau) and the free atmosphere above the plains which stretch out before the mountains. As the crow flies it is about 130 km between the 2 stations. For this comparison we will consider the soundings made at Munich on the 4th and 5th of September 1941 at 1800 hours and on the 5th and 6th of September at 0700 hours. Nearly simultaneously there were radiosoundings at Sulzau on the 4th at 1830 and on the 5th and 6th at 0630. The fourth datum was very easily interpolated using the Sulzau isotherms (Fig. 5). Using the simultaneous observations it was then possible to determine the averages for the two stations relative to diverse altitudes.

The first result, obtained by subtraction, concerns the "diurnal variation" of temperature at diverse heights above Munich and above Sulzau, drawn as curves in Fig. 10a. The diurnal variation is evaluated by the temperature difference between 1800 and 0700 hours. These values are acceptable, given that, starting from a certain height above the ground, the 1800 and 0700 observations give the approximate extremes of diurnal temperature variation in the free atmosphere, as known for the plains region from the Lindenberg soundings [7] and demonstrated for Sulzau according to what has previously been said.

Together with the evidence of direct temperature measurements for support, there is confirmation of the results obtained by Wagner [9] and Stapf [10], which were deduced indirectly from the diurnal atmospheric pressure at the different stations located on Alpine summits and which served as the basis for the new theory of valley winds: in the entire region of the valley atmosphere, the diurnal temperature variation is appreciably greater than the variation in the free atmosphere above the plain. At 1000 m altitude, the amplitude above Sulzau is <u>four times</u> that above Munich! Moreover, the amplitude at Sulzau decreases much more slowly as a function of altitude than at Munich. It remains almost constant in the first 500 meters, as

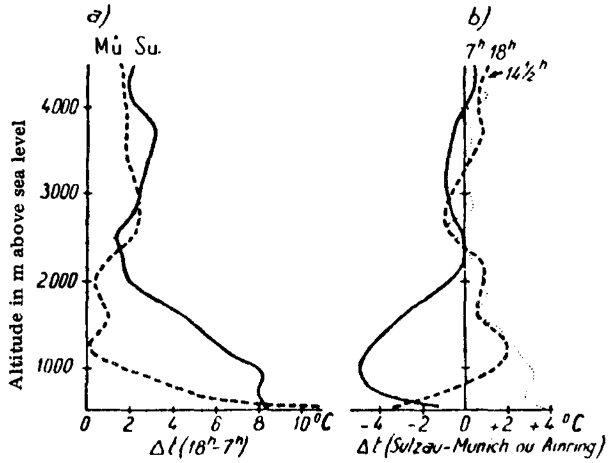

Figure 10. Simultaneous comparison of the temperature in the free mountain atmosphere and above the plain.
a) Diurnal temperature variation above Sulzau and above Munich (average of two days).
b) Temperature difference between Sulzau and Munich at 0700 and 1800 hours (German Summer Time) and between Sulzau and Ainring at 1430 hours.

already mentioned (see p. 81). It is only at 2400 m, the highest altitude of the mountains near Sulzau, that the two curves meet and remain very close to each other, varying around 2°C. This result is absolutely consistent with that obtained by the already mentioned refraction measurements of Brocks. For information, we present here a table of the diurnal variation of the indirectly determined temperature gradient.

Brocks concluded that the diurnal variation of density stratification decreases more slowly with altitude in the mountains than above the plain. Above the Central Alps, the diurnal variation is three times stronger than above the plain (*loc. cit.*, p. 72).

On the other hand, the temperatures of the two stations have been compared for both dates under consideration. Figure 10b shows the differences as a function of altitude.

Here again we find a well known fact: the morning valley air is colder (the difference can reach 5°C) and the afternoon valley air is warmer than the free atmosphere at the same level above the Bavarian plain. Only the low layers up to around 300 m are exceptions. In these near-ground layers the valley air is colder than above the plain even at 1800 hours. However, this is certainly exact only at the moment of formation of the cold near-ground layer which we have already mentioned several times. The radio sounding of September 5, 1941 at 1430 hours at Ainring, 5 km WNW of Salzburg airport (kindly reported to me by Prof. W. Georgii) provides the most striking confirmation of this. The dotted curve

Table: Diurnal variation of the thermal stratification above the plain (A) and above the Central Alps (B) (according to Brocks [6])

		Mean altitude (m)	Diurnal amplitude (°C/100 m)	Ratio of amplitudes
A	Karkopf-Surheim[1]	1080	0.38	0.21
	Karkopf-Hochkoenig	2340	0.08	
B	Hoh.Sonnblick-Taxenbach[2]	2100	0.50	0.64
	Hoh.Sonnblick-Hochkoenig	3020	0.32	

[1] Surheim, 430 m - Locality of the Upper Bavarian Plain, around 10 km from Salzburg down-valley from Salzach.
[2] Taxenbach, Farm above the locality, at 100 m - in the Salzach valley opposite the entry of the Rauris valley leading to Sonnblick.

(Figure 10b) represents the vertical distribution of the temperature difference between Sulzau and Ainring for the period under consideration. All of the valley, from the ground up to crest height, is therefore warmer than the plain in the afternoon.

The three curves of Figure 10b indicate that the height of the thermal equalization surface between the plains and the mountains is at 2400 m for the hours of observation. Comparison of the three stations (Munich, Ainring and Sulzau), seems to indicate that the temperature gradient between the mountains and the plain (which determines the valley wind circulation) is localized in the bed of the valley itself, i.e. it becomes effective only at the immediate border of the Alps. (Ainring is situated at the western border of Salzburg Basin). At least, since the Sulzau-Munich and Sulzau-Ainring temperature difference curves have the same appearance starting from around 1300 m, it seems likely that no essential temperature differences exist between Munich and Ainring.

The temperature differences between mountains and plain observed above the altitude of the thermal equalization surface are not sufficiently convincing, at least in the morning, that they alone could justify a definite conclusion in the sense of Wagner's notions [11] on the limiting surfaces of the thermal circulation. According to that hypothesis, a warm (cold) region of pressure compensation necessarily exists above a cold (warm) region. It is only at 1800 hours that the temperature of the 2400-3300-m layers above the plain is greater than that of the mountains, at which time the condition formulated previously would therefore be satisfied and would also agree with the value already found ([2, p. 230 ff]) for the upper limit of the "anti down-valley wind" above Sulzau.

In any case, it is already a very pleasing result that some soundings performed by quite different methods have indicated temperature differences between the plain and the mountain not only in the expected direction, but also at an acceptable order of magnitude.

The vertical temperature difference above Sulzau and Munich permits us directly to determine the pressure gradient (for dry air) at different altitudes. The result, presented graphically in Figure 11, can be expressed as follows: In the morning, the prevailing pressure gradient is oriented from the mountains towards the plain and attains its maximum value of almost 4 mbar/100 km (but with the reservation expressed in a remark made later on) right at the ground. It is consistent with the theory that the wind maximum, due to friction, is situated not at the ground itself, but a few hm above the ground--in our case 200 to 300 m above ground level. And as the theory requires, the horizontal pressure

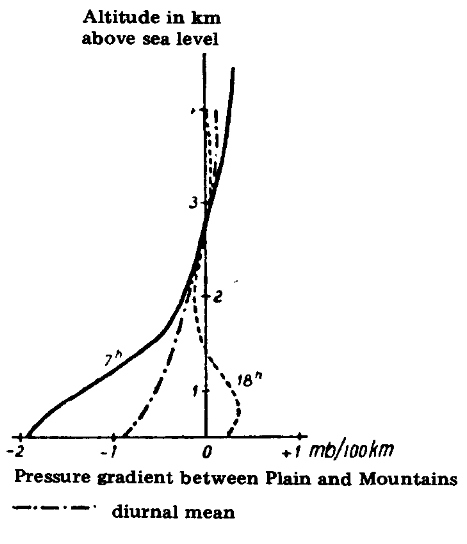

Figure 11: Pressure gradient between the plain and the mountains (Munich-Sulzau) at 0700 and 1800 hours (German Summer Time) and diurnal mean (0700 + 1800 hours)/2.

gradient decreases continuously with distance above the soil, becoming zero at 2800 m, where it then inverts and thus determines the upper compensating wind or "anti down-valley wind", as I have indicated [2, p. 229]. In contrast, the isobaric surfaces at 1800 hours are inclined towards the mountains (maximum pressure gradient 1/4 mbar per 100 km at 1000 m above sea level), but only up to around 1500 m. An inversion exists at this altitude, followed by still another inversion at 2750 m, where the absolute values reach their minima.

Considering the basic pressure values, which are probably not very precise (at Sulzau, for example, readings on the aneroid barometer, compared from time to time with hypsometric readings), it would seem more logical first to assume conditions which are symmetric relative to the diurnal mean of the two observations and to relate the pressure gradients at different altitudes to the diurnal mean. By taking 0700 + 1800 hours/2, we form an approximate diurnal mean--the dot-dash line in Figure 11--and obtain the diurnal pressure gradients as differences between the diurnal mean curve and the two plotted curves, which represent the times of maximum development of the up- and down-valley wind systems. We assume that the positive sign indicates that the pressure gradient is directed from the mountains towards the plain at 0700 hours and from the plain towards the mountains at 1800 hours.

Height Above Sea Level	Pressure Gradient
530-Ground	1.06 mbar/100 km
1000	0.80
1500	0.26
2000	0.01
2500	+0.00
3000	-0.00
3500	-0.06
4000	-0.015

The surface of barometric equalization would therefore be found at 2750 m, i.e. around 600 m above the mean crest altitude near Sulzau and 400 m above the thermal equalization level.

Temperature Differences Between the Air of the Valley and of the Summits

Now that we have observed the thermal contrasts between the atmosphere above the plain and that of the mountains, we can also consider the temperature differences, found in a series of measurements, between the air of the valley and of the summits. For such a comparison, we can use temperature recordings of two summit stations which are not too distant from each other[1]: Predigtstuhl, a 1578 m mountain in the Lattengebirge near Bad Reichenhall on the northern border of the Alps, and the Sonnblick Observatory (3106 m), situated on the central crest (Hohe Tauern). The first station is barely 30 km NW of our radio sounding station at Sulzau, and Sonnblick is 75 km SSW of Sulzau. Moreover, temperature measurements with an Assmann psychrometer were made every 15 minutes at the post on the Schlenken summit at 1649 m, situated nearby at 20 km down the valley. Our radio soundings permitted us directly to interpolate the temperatures corresponding to the altitude of the three summits above Sulzau and to compare them with the temperatures observed simultaneously at the mountain stations. The different values connected graphically relative to the differences between the summits and the free atmosphere above Sulzau represent the time evolution of the altitudes around 1580 m, 1650 m and 3100 m (see Figure 12). Despite the different distances of the three mountain stations from our aerologic

[1] Made available by the old Reichsamt fuer Wetterdienst at Berlin and the Sonnblick-Verein (Prof. H. V. Ficker).

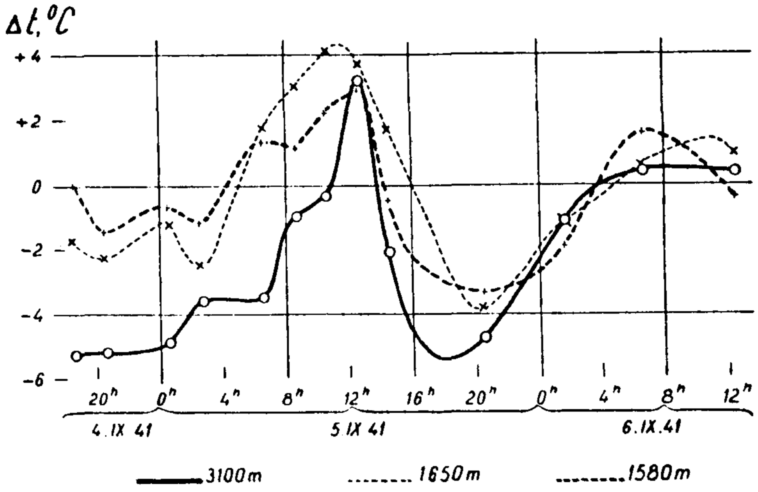

Figure 12. Comparison of the diurnal temperature variation of summit stations (Predigtstuhl 1578 m, Schlenken 1649 m and Sonnblick 3106 m) with those of the valley atmosphere at the same altitude (above Sulzau) (Δ = summit - valley air).

station, and despite the fact that their situations are different relative to the mountains (border-interior), we can consider--according to [12]--their temperatures to be representative of the temperatures of any summits of corresponding altitude situated in the region under consideration.

It is a repeatedly observed fact that the summits of mountains are generally colder than the free atmosphere, although at noon, particularly in clear weather, the mountains exhibit higher temperatures. However, the research work in this study has always dealt with comparisons with soundings made beyond the mountains (Friedrichshafen, Munich, Vienna, Lindenberg). Given that the mountain air at sufficient distance from the slope is warmer by day and colder by night than the free atmosphere above the plain, it remains to be seen in which direction and on which order of magnitude are the temperature differences between the summits and the free atmosphere of the mountains.

The profile of the curves in Figure 12 exhibits two characteristics: (1) In good weather, there may be positive and negative differences, where the mountains preferentially exhibit a lower temperature than the free mountain atmosphere during the night and a higher temperature during the afternoon (and probably from the morning in the low layers). This seems to favor the hypothesis of radiation effects. On the daily average, notably at the highest altitudes, the mountains are colder than the valley air. (2) The greater heating of the summits at noon seems to decrease with altitude, whereas the nocturnal cooling seems to increase. This result was already deduced from a comparison of the mean temperature differences between the Northern Alps and the free atmosphere above Munich [12] for higher altitudes (above 1600 m). That result was explained by the increasing influence of dynamic cooling of the currents crossing the mountains.

The above comparison also shows how, in particular cases, the temperature differences between the mountain summits and the free atmosphere above the valley can be large. During the two nights, the temperature drop at Sonnblick reached 5°C! We can assume that slopes also exhibit temperature differences which are more accentuated relative to the layers of free atmosphere at the same altitude. In any case, this immediately explains the energy reserves for the thermodynamic mechanism of slope wind circulation.

The considerations which we have just discussed permit us in conclusion to devote a few words to the structure of the atmosphere above a mountain range.

The Structure of the Atmosphere Above a Mountain Range

When we speak of temperatures at different altitudes in the mountains, our basis is measurements made at fixed mountain stations or at relatively short distances from the slopes. For example, for measurements made from cable cars, the term Slope Atmosphere would be more exact than the so-called Mountain Atmosphere to describe this "system of atmospheric particles lying side by side or superposed on each other along slopes, and whose formation depends on the conformation of the terrain, the state of the ground, exposure, insolation and radiation conditions, atmospheric currents, etc." [13, p. E1]. According to Geiger [14] , the action of different mountain atmospheres determines the so-called Mountain Atmosphere.

However, I would like to give the notion of "Mountain Atmosphere" rather the significance of an environment which is intermediate between the slope atmosphere adhering to the ground relief (a thin layer of atmosphere if we consider it from the macroclimatological viewpoint, measuring a few hm at most)--or the valley atmosphere, which has still to be defined below--and the atmosphere beyond the zone of influence of the mountain mass, qualified simply as Free Atmosphere. Without wishing by this notion to fix a precise limit between the Slope Atmosphere and the Mountain Atmosphere--and no more so between the other parts of the Atmosphere in the interior of and above the mountains--such a differentiation seems to me, according to experiments which have been performed, more appropriate for expressing the independence of different regions than the summation adopted by Geiger.

The Slope Atmosphere not only is a well defined region of the Mountain Atmosphere from the thermal viewpoint, expressing the direct influences of the ground[1] simply in the diurnal temperature variation, but it is also characterized from the dynamic viewpoint as the zone of up- and downslope winds, including the slope winds along the valley floor [15], glacier winds, etc.

Although the slope atmosphere appears essentially to be the layer adhering to the forms of relief of the mountains, the specific thermal structure which we have just studied, as well as the distribution of turbulence in the valleys and--we can readily generalize--in the hollows and depressions, require that we also regard this atmosphere which penetrates and occupies the concavities as an independent environment, designated by Valley Atmosphere. It is simultaneously the zone of an independent periodic circulation system, which is genetically well differentiated from the local currents along mountain slopes, i.e. the lower system of the valley wind circulation. We must also stress the fact that the valley atmosphere is a notion which applies only to large-scale

[1] Brocks, for example, on the basis of his optical measurements (*loc. cit.*, p. 69) notes: "The independence of the Slope Atmosphere is characterized in such a way that slopes 26 km 3 [*sic* -- 26.3 km ?] from each other exhibit similar effects as regards the neighboring field of atmospheric Solenoids".

depressions in the Mountainous Massif. It does not apply to small deformations in the ground relief which influence only the microclimate (e.g. crevasses along a slope, etc.). The atmospheric particles of such deformations constitute the elements of the slope atmosphere, which elements assume all the irregularities of the ground. In many cases the small depressions do not exhibit the characteristics of the valley atmosphere as just defined. On the contrary, this atmosphere is part of the "Slope Atmosphere". This is particularly true in the case where, because of the narrow cross section of such valleys, actual valley wind in the sense of Wagner's theory does not develop. Instead, transversal circulation develops because of fusion of the winds of the two valley slopes or--notably in the case of a steeply rising valley floor--a slope wind along the valley floor develops. The characteristic diurnal slope variation of the slope atmosphere, which is essentially influenced by the ground relief, will extend over the entire cross section of the valley.

By Mountain Atmosphere, we then designate that part of the atmosphere located between the slope atmosphere or valley atmosphere and the unperturbed atmosphere at great distance (horizontal and vertical) from the mountain chain.

As its name indicates, the mountain atmosphere carries the imprint of thermal and dynamic influences marked by the Mountainous Massif in its entirety. A completely independent diurnal temperature variation develops in this atmosphere. The circulation of this atmosphere exists in the form of a compensating current for the overall circulation of mountain winds. Its radius of action is ultimately determined by the deformation field of the current perturbed by the obstacle.

We thus arrive at a classification of the atmosphere based for the time being mainly on the diurnal temperature variations and on the state of atmospheric circulation in the region of a mountain range. This type of classification is represented schematically in Figure 13.

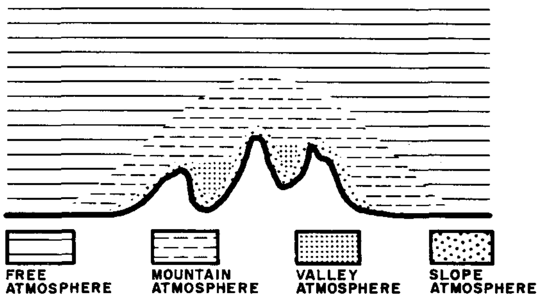

Figure 13: Diagram of the structure of the atmosphere above a mountain range.

It will be an extremely interesting and profitable task for Alpine meteorological research to expand on those notions on the structure of the mountain atmosphere which have served as our working hypothesis, and in particular to find mathematical bases for a clearer separation of the different regions and to analyze their thermal and dynamic structure in detail. To achieve this goal, it would be necessary to make more use of modern aerological resources, particularly powered flight and gliding. The direct measurements of atmospheric structure above well defined regions in space could be advantageously supplemented by the indirect method of ground refraction observations used successfully by Brocks. This method is interesting in that it yields values for extended atmospheric layers and thus compensates for differences of accidental nature.

[Ed. Note: Dr. Ekart's paper concludes with a photograph captioned "the mountain atmosphere". Unfortunately, the original photograph was not available and it could not be properly reproduced here.]

REFERENCES

1. A. Wagner. - Theorie und Beobachtung der periodischen Gebirgswinde [Theory and observation of periodic mountain winds]. Gerl. Beitr. Geoph., 52, 408, 1938.

2. E. Ekhart. - Beitraege zur alpinen Meteorologie [Contributions to Alpine meteorology]. Met. Z. 1944, 217.

3. A. Huber. - Temperaturanstiege und-maxima bei Nacht als regelmaessige Erscheinungen des Temperaturverlaufes [Temperature increases and maxima at night as regular features of the temperature variation]. D. Met. Jahrb. Bayern 1926, Anh. C.

4. W. Schmidt. - Der Massenaustausch in freier Luft und verwandte Erscheinungen [Mass exchange in the free atmosphere and related phenomena]. Probl. kosm. Phys. VII, Hamburg 1925.

5. H. v. Ficker. - Die Flaechen gleichen Druckes bei Berg und Talwind [Isobaric surfaces in mountain and valley wind]. Taet.-Ber. Pr. Met. Inst. Berlin 1920-23, 56.

6. K. Brocks. - Lokale Unterschiede und zeitliche Aenderungen der Dichteschichtung in der Gebirgsatmosphaere. (Beitraege zur Untersuchung der periodischen Gebirgswinde). [Local differences and time changes of the density stratification in the mountain atmosphere (Contributions to the study of the periodic mountain wind)]. Met. Z. 1940, 62.

7. H. Hergesell. - Der taegliche Gang der Temperatur in der freien Atmosphaere ueber Lindenberg [The diurnal temperature variation in the free atmosphere above Lindenberg]. - Arb. Pr. Aeron. Obs. Lindenberg 14, 1922.

8. F. M. Exner. - Abhaengigkeit des Turbulenzfaktors der Winde von der vertikalen Temperaturverteilung [Dependence of the wind turbulence factor on the vertical temperature distribution]. - Sitz. Ber. Akad. Wiss. Wien. 136, 452, 1927.

9. A. Wagner. - Der taegliche Luftdruck- und Temperaturgang in der freien Atmosphaere und in Gebirgstaelern [The diurnal air-pressure and temperature variation in the free atmosphere and in mountain valleys]. Gerl. Beitr. Geoph. 37, 315, 1932.

10. H. Stapf. - Der taegliche Luftdruck und Temperaturgang in der freien Atmosphaere in seiner Abhaengigkeit von den Jahreszeiten [The diurnal air-pressure and temperature variation in the free atmosphere as a function of the season]. - Diss. Berlin 1934.

11. A. Wagner. - Über die vertikale Maechtigkeit einer thermischen Zirkulation und die vertikale Temperaturschichtung in derselben [The vertical thickness of a thermal circulation and the vertical temperature stratification in that circulation]. - Gerl. Beitr. Geoph. 51, 410, 1937.

12. E. Ekhart. - Mittlere Temperaturverhaeltnisse der Alpen und der freien Atmosphaere ueber dem Alpenvorland [Mean temperature conditions of the Alps and of the free atmosphere above the Alpine foothills]. - Met. Z. 1939, 12, 49.

13. A. Buedel. - Die Zugspitzbahnversuche III Teil [Experiments on the Zugspitze telepherique]. - D. Met. Jahrb. Bayern 1931, Anh. E.

14. R. Geiger. - Das Klima der bodennahen Luftschicht [The climate near the ground]. - 2. Aufl. 1942.

15. E. Ekhart. - Neuere Untersuchungen zur Aerologie der Talwinde. Die periodischen Tageswinde in einem Quertale der Alpen [New studies on the aerology of valley winds. The periodic diurnal winds in an Alpine cross valley]. - Beitr. Phys. fr. Atm. 21, 245, 1934.

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Friedrich Defant

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Friedrich Defant

Summary

This work is concerned with mountain winds, especially with periodic diurnal slope winds. A short review of A. Wagner's theory of mountain winds and their periodic diurnal variations is presented, as well as a summary of present knowledge of the slope winds on mountain slopes. Mean profiles of wind velocities above slopes are determined for the upslope wind and the downslope wind for undisturbed normal days, using observations and pilot balloon measurements. Then, a new theoretical treatment of stationary slope currents developed by L. Prandtl is thoroughly discussed. The results of this theory (i.e. theoretical profiles of wind velocities and temperature anomalies for the upslope and downslope winds) are compared in detail with the observed data. The agreement between theory and observation is found to be surprisingly good in the air layers of interest (i.e. up to an altitude of 100 m, measured normal to the slope). Prandtl's theory is also extended to include periodic diurnal variations of slope winds: it turns out that the solutions for wind velocity and temperature anomaly are to be multiplied by the periodic factor cos σ, where σ is the diurnal period. It is also shown that changing the value of ν (coefficient of friction) or the value of a (coefficient of thermal conductivity) has little effect on the whole phenomenon.

The solution for the velocity of slope winds is discussed and compared with the solution obtained by A. Defant for heavy air masses flowing down inclined ground; the two theories are in satisfactory agreement on points for which they can be compared. The question of which value should be chosen for the coefficient of friction of forest-covered, irregular slope surfaces is discussed in detail, along with the use of the coefficient of friction in calculations. Finally, mean approximate calculations are made of the air transport of the slope-wind circulation and its effect on the generation of mountain and valley winds, using the vertical [Translator's note: In this article the author often uses "vertical" to mean "normal to the slope" rather than "parallel to g". In some sentences we have retained the word "vertical" to avoid cumbersome phrasing; thus "vertical distribution of velocities above the slope" could be more accurately rendered as "distribution, as a function of the space coordinate normal to the slope, of the velocity component parallel to the slope."] distribution of velocities above the slope; useful mean results are obtained, and they afford an insight into the mechanism of slope winds.

I. Introductory remarks on A. Wagner's views on mountain winds

In recent years the problem of periodic mountain winds has been treated in numerous articles by A. Wagner, E. Ekhart, and other authors, and has been the subject of a comprehensive theory by A. Wagner. Wagner made a decisive step forward in that he distinguished between slope winds, mountain winds, valley winds, and equalizing currents between lowlands and tablelands. His theory satisfactorily resolved the earlier differing conceptions of mountain winds and their origin, supplying a more rigorous foundation [1-31].

Practical research on mountain winds and their mechanism has been greatly advanced by the numerous investigations of E. Ekhart, A. Jelinek, E. Moll, A. Riedel, and others. This research has proceeded in parallel with the evolving views of A. Wagner [1-31].

In 1938 E. Ekhart, in an easily understandable article [30] titled "The Diurnal Winds of the Alps", treated the whole system of mountain winds and discussed questions relating to air circulation between plain and mountains, whose mechanism he had investigated with A. Burger in an earlier article [25].

It is hoped that it will suffice here merely to mention this comprehensive series of articles, since the present work deals with only one part of the mountain wind problem, namely the periodic diurnal variation of slope winds. I take the liberty of adopting the basic train of thought of A. Wagner's theory [31] at the beginning of this investigation, and then establishing its special connection with periodic diurnal slope winds.

Mountain slopes become progressively warmer after sunrise, and air begins to flow upward along the slopes in the form of upslope winds. [Translator's note: The description of the diurnal course of slope and valleys winds which begins here is illustrated in Figures 1 through 8.] The valley air as a whole is colder at this time than is the air over the plain. Therefore the pressure drop is in the direction valley-to-plain, and the down-valley wind is still in progress. (This is assuming that a pronounced pressure drop is not present at the altitude of the valley crests.) At the altitude of the crests (or higher), the air ascending along the slopes bends back toward the middle of the valley. Then it flows downward and is carried by the down-valley wind (in all layers, down to the ground) toward the exit of the valley. Therefore the cold (relative to the plain) valley air as a whole is gradually replaced by the warmer slope air. As a consequence of this, the valley air and the air over the plain will have the same average temperature at some point in time. The isobaric surfaces are then horizontal, a pressure drop is no longer present, and the down-valley wind ceases. This is the null point between the down-valley wind and the up-valley wind, and at this time the slope wind circulation follows the cross-section of the valley. All the air which ascends along the slopes descends over the middle of the valley and returns to the slopes. This is also the point at which the valley air is most rapidly warmed.

As the valley air is further warmed with respect to the air over the plain a plain-to-valley pressure drop develops, and the up-valley wind begins, increasing in intensity. The more air which the up-valley wind carries into the valley, the less is the downward motion of the circulation of slope air in the middle of the valley, and the increasing volume of air which is ascending along the slopes is drawn more and more into the upper compensation current of the up-valley wind.

As the afternoon continues, the upslope wind gradually dies out as direct warming of the slopes decreases. Almost all of the (continually decreasing) air flow of the upslope wind is now carried above the altitude of the crest by the up-valley wind of the plain. The valley air attains its maximum temperature; the plain-to-valley pressure gradient is at its highest value, and when the slope wind circulation is completely extinguished in late afternoon the air merely flows parallel to

(a)

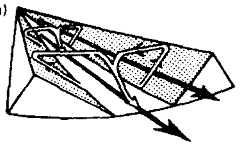

Figure 1: Air flow in the morning at sunrise. The upslope winds begin (white arrows). The mountain wind (black arrows) is still present in the valley. The pressure drop is in the down-valley direction. Temperature: valley is cold, plain is relatively warm. Changes in temperature (until situation shown in Figure 2): valley is becoming warmer, plain is becoming cooler.

(b)

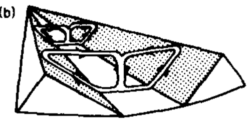

Figure 2: Air flow in the forenoon (about 9 a.m.). The upslope winds are strong. The system is in a state of transition from mountain wind to valley wind. Pressure drop: zero. Temperature: equalized. Changes in temperature (until the situation shown in Figure 3): valley is rapidly becoming warmer, temperature over the plain is changing slightly.

(c)

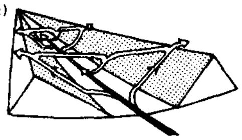

Figure 3: Air flow from noon to early afternoon. The upslope winds are decreasing. The valley wind is fully developed. The pressure drop is in the up-valley direction. Temperature: valley is warm, plain is relatively cold. Changes in temperature (until the situation shown in Figure 4): essentially none.

(d)

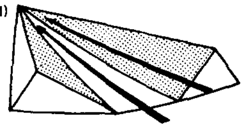

Figure 4: Air flow in the late afternoon. The upslope winds have died out. The strength of the valley wind is the same as in Figure 3. The valley wind extends to the sides of the slopes. The pressure drop is in the up-valley direction. Temperature: valley is warm, plain is rela- ely cold. Changes in temperature (until the situation shown in Figure 5): valley is slowly cooling.

(e)

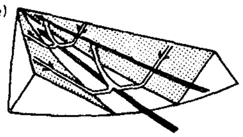

Figure 5: Air flow in the evening. The downslope winds are beginning. The valley wind is decreasing. The pressure drop is still in the up-valley direction. Temperature: it is slightly warmer in the valley than it is over the plain. Changes in temperature (until the situation shown in Figure 6) the valley is cooling rapidly; the plain is cooling only slightly.

(f)

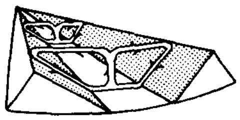

Figure 6: Air flow from late evening through the first half of the night. Downslope winds are present. The system is in a state of transition from valley wind to mountain wind. Pressure drop: zero. Temperature (until the situation shown in Figure 7): valley continues to cool rapidly.

(g)

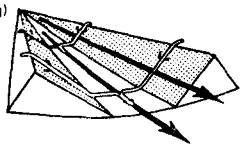

Figure 7: Air flow at night. Mountain wind is present along with the downslope winds. The pressure drop is in the down-valley direction. Temperature: valley is cold, plain is relatively warm. Changes in temperature (until the situation shown in Figure 8): valley is cooling; plain is cooling slightly but is warm relative to the valley.

(h)

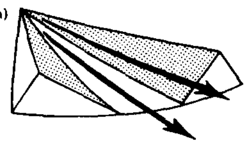

Figure 8: Air flow from night until morning. The downslope winds have died out. The mountain wind extends to the sides of the slopes. The pressure drop is in the down-valley direction. Temperature: cold in the valley, warmer over the plain. Changes in temperature (until the situation shown in Figure 1): there are only slight changes in the valley and over the plain.

Figures 1 through 8: Schematic representation of the air flow in a valley during a normal day the valley floor. This flow (the up-valley wind) extends to the sides of the mountain slopes.

Then the valley air begins to cool slowly. This cooling process is increasingly aided by the downslope winds which begin at this point. The upward motion over the middle of the valley likewise has a continual cooling effect on the valley air as a whole, and this cooling effect is aided by the emission of radiation now beginning.

Thus the moment arrives at which the temperature of the valley air is equal to the temperature of the air over the plain. The isobaric surfaces become flat, and the up-valley wind passes through the null point and is converted into the down-valley wind.

The slope wind circulation again takes place only in the valley cross section; the air which flows down the slopes rises over the middle of the valley and flows back to the slope at the altitude of the crest. This corresponds to the situation shown in Figure 2 except that the directions of flow are reversed. The down-valley wind now begins, and an increasing fraction of the downslope air flow is carried to the exit of the valley by the down-valley wind.

The valley air as a whole attains its nighttime temperature minimum; the upward motion over the middle of the valley ceases entirely, and the downslope winds finally cease. The air flow is again parallel to the valley floor, and because of the absence of slope winds the down-valley wind extends to the sides of the mountain slopes in a manner similar to that shown in Figure 4, but with reversed flow direction.

At sunrise the slopes begin to be warmed and the upslope wind begins again. This closes the cycle of this periodic diurnal circulation system; a situation like the one shown in Figure 1 again ensues.

In the interplay between the mountain wind, valley wind, and slope winds, the phase of development of the up-valley and down-valley winds is displaced. The whole picture of wind circulation in the valley depends on whether this displacement is large, small, or equalized. It is not necessary here to consider the seasonal variation of the winds in detail because this variation is quite understandable if one considers (with A. Wagner) the different course of the temperature variation in the valley and in the plain, as well as the different irradiation of the slopes in summer and winter (especially in the case of a snow cover in the latter part of the year).

We now direct our attention to slope winds, the structure of which has already been studied during aerological investigations conducted by the Innsbruck Meteorological Institute (although data was taken on only a few days under different external conditions). The daily development of slope winds of certain intensities and vertical extents is understandable merely from the viewpoint of daily warming and cooling caused by absorption and emission of radiation. It is also understandable that an undisturbed circulation motion (in the sense of Wenger's slope wind theory based on the Bjerknes circulation theorem) occurs only when undisturbed conditions of absorption and emission of radiation are present and the surface properties of the slope approach ideal uniform conditions. Each surface anomaly, whether it be afforestation or furrows and ditches, has a modifying effect and impresses on the slope winds a more turbulent character, deviating from a uniform state. Shading of the slopes also plays an important role in the uniform development of slope winds.

The decisive factors for slope winds, however, are the temperature conditions on the slope itself (or in its immediate neighborhood) and the temperature conditions at some distance from the slope (say roughly above the middle of the valley at the same altitude). According to Wenger, the diurnal temperature variation at a point on the slope and the diurnal temperature variation at a corresponding point at the same altitude in the open valley constitute (because of their differing magnitudes) a decisive driving force which maintains the slope wind circulation. Jelinek's observations that a large superadiabatic temperature gradient is present in the case of upslope winds, that a ground inversion is present in the case of downslope winds, and that isothermal conditions are present at the time of wind reversal are very informative. Likewise, the fact that slope winds change direction (from upslope to downslope) when the temperature gradient changes (from increasing with increasing altitude to decreasing with increasing altitude) is an insight which has been gained by consideration of the data.

The layer thickness of the slope wind (measured normal to the slope) is about 200 m according to present knowledge. However, this is only a rough value, and the layer thickness can reach about 260 m under undisturbed fair-weather conditions with good development of slope wind circulation. The nocturnal downslope wind apparently has smaller layer thicknesses, which is understandable. The layer thickness seems to increase along the slope profile in the direction of the crest altitude. Therefore the upslope wind layer can be pictured as a wedge-shaped (thicker at higher altitude) wind layer lying on the slope.

The data indicate that the component of velocity parallel to the slope is of the order of magnitude of 2-4 m/sec, but this applies only for an average slope inclination of about 30° and changes very sharply with the angle of inclination of the slope. In the immediate neighborhood of the surface the parallel component and the perpendicular component of velocity both seem to have small values, and their maximum values are concentrated more at the middle altitudes of the slope wind layer.

Relatively speaking, the slope wind layer is very thin, and thus it is understandable that it reacts quite rapidly to any external influence: Any change in incident radiation causes the slope wind flow to react immediately. Thus, for example, the slope wind circulation is sensitive to disturbances caused by transient shading of the slope, and the wind change from downslope to upslope (and conversely) is related to the migration of the shade boundary along the slope. Exposure also plays an important role in the development of the slope wind flow; exposure is related to the initiation, intensity, and cessation of slope winds.

This short introduction is sufficient for the purposes of this article. My father (Dr. A. Defant) has called my attention to a fairly recent theoretical explanation of the formation of slope winds due to L. Prandtl [33, p. 373]. This derivation apparently permits a deeper insight into the nature of slope winds.

Following this suggestion, I have devoted much attention to Prandtl's model, which has not been mentioned in meteorological circles up to now. In particular, I have considered the relation of this work to my father's investigation "The Flow of Heavy Air Masses Down Slopes", with which it exhibits some noteworthy connections which seem sufficient for a detailed treatment [32].

Prandtl treats the flow of stratified air on slopes in a section titled "Mountain and Valley Winds in Stratified Air", written from the viewpoint that previous treatments are incorrect. In order to avoid vagueness, Prandtl limits himself to a treatment of slope winds alone. For the exposition which follows here, however, it is essential to provide (using the available measurements and pilot balloon data) a mean quantitative picture of upslope and downslope winds, even if this picture is more or less crude; this is done in the following section.

II. Derivation of mean vertical velocity profiles of upslope and downslope winds, using data from pilot balloons on mountain slopes

The purpose of all the observations (including pilot balloon tests and temperature measurements) which have been made up to now on the mountain slopes of the Inn Valley was to gain a deeper insight into the nature of slope winds and to demonstrate the role which slope winds play in the mountain wind circulation, considered from a purely rational viewpoint.

Wenger's [2] brief, purely schematic, rough calculations are not satisfactory, and they leave open the question of the vertical distribution of velocities (i.e. the distribution, as a function of altitude, of the velocity component parallel to the slope). Also, the measurements which have been made up to now are generally limited to determination of the horizontal and vertical components of the velocity of the pilot balloon at each point of its path, along with determination of the altitude of the pilot balloon relative to the point at which it was released.

This treatment of slope winds, however, does not seem to be very suitable because slope winds, as their name indicates, are currents which depend on (and are in fact caused by) the inclination of the slope. Thus it is not completely satisfactory to disregard the inclination of the slope by using a coordinate system whose origin is at the point of release of the pilot balloon and which has a horizontal x-axis and a vertical z-axis. A coordinate system is needed whose origin is at the point of release but which has its x-axis in the slope surface and perpendicular to the mountain ridge, and its z-axis perpendicular to the slope surface. Therefore we abandon (as does Prandtl) the horizontal and vertical coordinates x and z, and define the "slope" coordinates s (in the plane of the slope) and n (normal to the slope). The treatment of the flow as a two-dimensional problem excludes having the third coordinate (in the plane of the slope) be parallel to the direction of the valley.

I used the measurements of A. Riedel [19], taken on the slope of the Innsbruck Nordkette, to construct mean profiles of the slope wind velocity component in the direction of the slope, for upslope and downslope winds. Also, for a total of 11 vertical profiles of pilot balloon ascents (double sighting), I transformed into the above-described slope coordinate system the published values of the altitudes (relative to the point of release) of the balloon at points along its path. The 11 ascents which were chosen were the ones made under upslope wind conditions which were as little disturbed as possible, with no noticeable external disturbing influences. The slope profile, which was more or less inclined at various points, was approximated by rectilinear pieces, and the horizontal and vertical components of the velocity at each point of the path were decomposed into components parallel to the slope profile. This was done as exactly as possible, using the various local inclinations of the slope. Likewise, the altitude (normal to the slope) was calculated for each point of the path.

Then, assuming stationary velocity conditions throughout all layers, I combined the path points at various distances from the point of release to form an average line normal to the slope -- as is customary in work with pilot balloon data. Thus I was able to obtain the distributions (normal to the slope) of the velocity components parallel to the slope.

Table 1 gives the average values (in m/sec) of these components of wind velocity parallel to the slope for altitude steps of 5 m (measured normal to the slope), up to an altitude of 130 m. The data upon which these values are based were taken during 11 pilot balloon ascents in upslope wind.

It is clear that only simultaneous, exact measurements at all these altitudes could yield true values. However, the reasonable distribution of velocity as a function of altitude indicates that even such a forced, crude calculation can yield usable values, provided that favorable ascent conditions are chosen and the evaluation is performed as carefully as possible.

Almost all of the 11 velocity distributions show an increase in velocity from the ground up to altitudes of 20 to 40 meters, and then a decrease in velocity. The values are, of course, rather different at higher altitudes. It must be understood that, soon after passing through the slope wind layers, the balloon is seized by the more or less pronounced gradient wind currents which exist above the slope wind layers, or seized by the actual mountain or valley winds, which then gradually affect the velocity by an increasing amount.

These upper layers apparently do not disturb the rather obvious pattern of velocity distribution in the slope wind layers, which doubtless obeys some regular principle. Since the pilot balloons were released at different times of day, there are variations in the speeds of the individual ascents. This fact in itself means that the data are suitable only for averaging. Figure 9 shows the mean curve.

We see that the velocity gradually increases from the ground up to an altitude of 27 m. At this altitude the velocity component parallel to the slope is maximum (3.9 m/sec). Then a gradual decrease in velocity begins. The velocity levels off at about 2.3 m/sec at altitudes of 110 to 130 m. This description is sufficient here; later I shall return to the other curves shown in Figure 9.

Essentially the same method was used in constructing a corresponding downslope wind profile. In this case also, I used only pilot balloon data which were taken under downslope wind conditions which were as undisturbed as possible, and determined (using a similar decomposition of components) the velocity components (parallel to the slope profile) and the altitude (normal to the slope) for each point of the path.

Table 2 gives the results (in m/sec) of averaging the data from 5 ascents.

In this case also, the data for each ascent shows an increase in velocity with increasing altitude, up to altitudes of 20 to 40 m. Then there is a decrease in velocity which becomes erratic at higher altitudes.

Table 1.
Velocity component w (in m/sec) parallel to the slope as a function of n. (Average values of data taken during 11 pilot balloon ascents in upslope wind.)

Altitude n (m)	Average Value (m/s)
5	2.29
10	2.91
15	3.39
20	3.70
25	3.86
30	3.86
35	3.80
40	3.67
45	3.53
50	3.39
55	3.24
60	3.09
65	3.04
70	2.89
75	2.76
80	2.76
85	2.61
90	2.53
95	2.48
100	2.42
105	2.35
110	2.29
115	2.27
120	2.26
125	2.29
130	2.38

Table 2.
Velocity component w (in m/sec) parallel to the slope as a function of n. (Average values of data taken during 5 pilot balloon ascents in downslope wind.)

Altitude n (m)	Average Value (m/s)
2.5	0.63
5	0.96
10	1.50
15	1.95
20	2.22
25	2.34
30	2.35
35	2.27
40	2.15
45	2.04
50	1.91
55	1.81
60	1.74
65	1.66
70	1.60
75	1.49
80	1.34
85	1.10
90	0.86
95	0.40
100	0.23
105	0.19
110	0.02

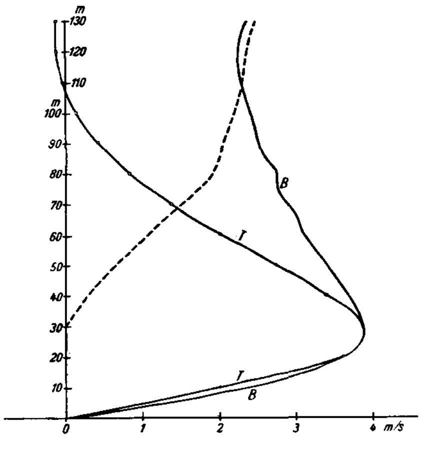

Figure 9. Profile (normal to the slope) of upslope wind velocity. B = observed values; T = theoretical values; broken curve = difference between T and B.

Averaging yields a velocity distribution which corresponds to the downslope wind flow. Figure 10 shows this average curve.

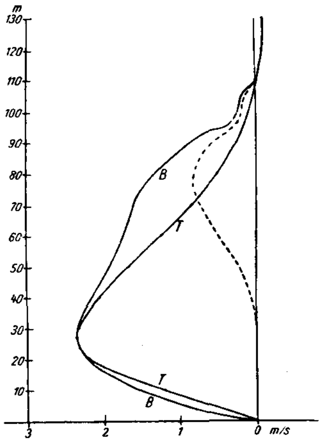

Figure 10. Profile (normal to the slope) of downslope wind velocity. B = observed values; T = theoretical values; broken curve = difference between T and B.

Again there is an increase in velocity from the ground up to an altitude of 27 m, with a maximum of 2.36 m/sec (which is lower than the maximum in the upslope case). Then there is a decrease in velocity with increasing altitude. No doubt the disturbing influences of the mountain and valley winds and/or the gradient wind are responsible for the appearance of the upper part of the curve.

To summarize: The observations show that there is an increase in wind flow parallel to the slope profile up to an altitude of 27 m, on the average, for both upslope and downslope winds. Understandably, the maximum value is higher in the case of the upslope wind (3.9 m/sec) than it is in the case of the downslope wind (2.36 m/sec). From this point upward, the curves show a decrease in velocity. On purely rational grounds, the wind velocity should decrease to zero at the upper boundary of the slope wind layer. The decrease is clearly recognizable in the averaged curves, but eventually the disturbing influence of the gradient wind flow and/or the mountain and valley winds is present, and this prevents the velocity from decreasing to zero at the upper boundary of the slope wind layer.

So much for an approximate treatment of the vertical profiles of velocity.

We now turn our attention to the thermal stratification normal to the slope. Unfortunately this has not been measured for high altitudes. The only temperature measurements which exist are those made on the intermediate towers of the Innsbruck Nordkette cable car and the Patscherkofel cable car up to altitudes of 20 m above the ground [27].

These temperature observations were made on days when the slope had a snow cover, which caused the very lowest layers (up to 2 m) to have an especially low temperature. Observations were also made on days when a true upslope wind was present. When an upslope wind was present there was a slight increase in temperature in the upper part of the layer (up to 20 m height). This increase in temperature gradually turned into the normal decrease of temperature with increasing altitude. The temperature gradient normal to the slope is undoubtedly disturbed by the snow cover. It is to be expected that, in normal cases without a snow cover, a superadiabatic gradient exists in the very lowest layers when an upslope wind is present. A. Wagner remarks that, according to measurements which he has made, this strong superadiabatic temperature gradient is a characteristic property of the upslope wind. Unfortunately he has not given a numerical treatment of this. In the case of the downslope wind, the snow cover naturally aids the low temperatures of the very lowest layers (which low temperatures in fact cause the downslope wind). The observations also show that a ground inversion is always present in the case of the downslope wind. This inversion turns into the normal decrease of temperature with altitude for altitudes higher than that of the slope wind maximum.

III. Prandtl's Theory of Stationary Slope Wind Currents

L. Prandtl [33, p. 373] treats the slope wind problem from the following viewpoint. Consider a slope which is inclined at an angle α with respect to the horizontal, and suppose that above this slope is a body of air with stable layers (i.e. the potential temperature of the air increases with increasing altitude). A small volume element of air which is embedded in a stratified air mass of this type can, upon being heated, rise only to the altitude at which it is surrounded by air of the same temperature. If, however, heat is transferred to the volume element of air from the base of the system (i.e. from the surface of the slope), then the volume element rises further, depending on the duration and intensity of the heat transfer.

If the temperature of the inclined surface of the slope is somewhat higher than the temperature of the air which is resting on it, then the volume elements of this layer of air will move with a uniform velocity (neglecting the build-up time at the lower edge of the slope).

As already mentioned, the coordinates x (horizontal) and z (vertical) are replaced by the coordinates s (parallel to the slope) and n (normal to the slope).

If we now assume that the potential temperature θ is a linearly increasing function of the form A + Bz, and add to it a certain temperature anomaly θ' caused by the conduction (in the n-direction) of heat from the heated surface of the slope, then we obtain an expression for the potential temperature in the form

$$\theta = A + Bz + \theta'(n), \tag{1}$$

In this expression the cause of motion of air up or down the slope is given by the vertical buoyant force $g\,\frac{\rho' - \rho}{\rho'}$ where ρ' is the unperturbed density and ρ is the density of the perturbed volume element under consideration (see 36, p. 47).

Now $\rho = \frac{\rho' T'}{T' + \Theta'}$, and as a first approximation we obtain $g\beta\theta'$ for the upward acceleration (with θ' positive), where we have set $\frac{1}{T'} = \beta$. The acceleration due to buoyancy thus takes place in the upslope direction.

The velocity w parallel to the slope in the s-direction is expected to be a function of n, and its time derivative must equal the sum of the buoyancy parameter $g\beta \sin\alpha \cdot \theta'$ and the friction parameter $\nu\, \partial^2 w/\partial n^2$:

$$\frac{dw}{dt} = g\,\beta \sin\alpha \cdot \Theta' + \nu \frac{\partial^2 w}{\partial n^2} \tag{2}$$

and for the stationary state (dw/dt = 0) we have

$$0 = g\,\beta \sin\alpha \cdot \theta' + \nu \frac{\partial^2 w}{\partial n^2} \tag{3}$$

The equation of heat conduction for the heated surface of the slope yields the relation

$$\frac{d\Theta}{dt} = a\left(\frac{\partial^2\Theta}{\partial n^2} + \frac{\partial^2\Theta}{\partial s^2}\right) \tag{4}$$

where a is the coefficient of thermal conductivity.

For air moving in the s-direction, we set

$$\frac{d\Theta}{dt} = \frac{\partial\Theta}{\partial t} + w\,\frac{\partial\Theta}{\partial s} \tag{5}$$

in which $\partial\theta/\partial t = 0$ for the stationary state. If we neglect $\partial^2\Theta/\partial s^2$ in comparison to $\partial^2\Theta/\partial n^2$ in (4), which is permissible, then we obtain the equation

$$w \cdot \frac{\partial\Theta}{\partial s} = a \frac{\partial^2\Theta}{\partial n^2} \ . \tag{5a}$$

Using the transformation of coordinates $z = s \cdot \sin\alpha + n \cdot \cos\alpha$, Equations (5a) and (1) yield

$$w \cdot B \sin\alpha = a \frac{\partial^2\Theta'}{\partial n^2} \ . \tag{6}$$

Differentiating twice with respect to n and inserting the result in Equation (3), we obtain

$$0 = g \sin\alpha\, \beta\, \Theta' + \frac{a\,\nu}{B \sin\alpha} \cdot \frac{\partial^4\Theta'}{\partial n^4} \ . \tag{7}$$

The simple solution of this differential equation is

$$\Theta' = C \cdot e^{\frac{-n}{\ell}} \cdot \cos\frac{n}{\ell} \ , \tag{8}$$

where

$$\ell = \sqrt[4]{\frac{4a\nu}{g\beta B \cdot \sin^2\alpha}} \ . \tag{9}$$

If we differentiate the solution (8) twice with respect to n and insert the result in (6), then we find the following equation for w:

$$w = C \cdot \sqrt{\frac{g\beta a}{B \cdot \nu}} \cdot e^{\frac{-n}{\ell}} \cdot \sin\frac{n}{\ell} \ . \tag{10}$$

Thus we have solved the problem for the case of a stationary flow along the slope, and obtained the temperature anomaly and the velocity w as functions of n (which is the coordinate normal to the slope).

Prandtl assumes a value for C in the expressions for the solutions (8) and (10), and gives a graphical representation of the curves for θ' and w. (C is a measure of the temperature anomaly at the surface of the slope, where n = 0.)

We see that, in the lower layers, w rapidly increases with increasing altitude normal to the slope (i.e. in the n-direction), reaches a maximum, and then approaches zero less rapidly than the previous increase to the maximum value; it even exhibits a small negative value at higher altitudes.

The curve of the potential temperature anomaly θ' has the assumed value C at the ground and then decreases with increasing altitude n, passes through zero, and has small negative values for altitudes above that. According to Prandtl the reason for these negative values is that, because of friction, layers which have not themselves been heated are moved by the thermal buoyancy effect and (after being raised to a higher altitude) are colder than the environment in the unperturbed state."

It is of interest to see how well the simple stationary form of Prandtl's theory can be made to fit the average data which I have given above.

IV. Discussion of Prandtl's theoretical solution, and comparison of it with the data

In order to compare Prandtl's theory with the data, it is necessary to discuss in detail the constants to be used in the calculations.

The assumption of a temperature anomaly of 5°C at the surface of the slope (n = 0) probably corresponds rather well to average conditions, so we set C = 5°C. The stratification of potential temperature is to be established at an average value of 0.4°/100 m, so $B = 4 \times 10^{-5}$ degree$\cdot$cm^{-1}. We assume the value of $\beta = 1/T$ to be

1/283°. We assume that the average angle α of inclination of the slope is 42.5°, which corresponds to the inclination of the ground at the site in the Innsbruck Nordkette (Seegrube) where the observations were made.

The value ν of the coefficient of friction and the value a of the coefficient of thermal conductivity will be treated in detail.

If A_τ is the exchange parameter due to transport of momentum (apparent friction) and A_q is the exchange parameter due to transport of heat content (apparent heat conduction), then $\nu = A_\tau/\rho$, and $a = A_q/\rho$, where ρ is the density of the air.

For turbulent flow conditions, simultaneous measurements of the velocity and temperature profiles in the wake of a heated rod have clearly indicated that the quantity $a/\nu = A_q/A_\tau$ has the value 2. On the other hand, measurements based on mixing an air jet with air jets of a different temperature have likewise yielded the value 2 (see [33, p. 108]). Up to now it has been customary, for turbulent frictional layers at a wall, to set $A_q = A_\tau$, i.e., $a/\nu = A_q/A_\tau = 1$. This value was always in agreement with experiment.

However, more recent measurements of the velocity and temperature profiles for plates and tubes performed by F. Elias and H. Lorenz in 1930 and 1934 gave a value A_q/A_τ = 1.4 to 1.5, which must be regarded as the best value available at present.

For conditions in the atmosphere, A_q and A_τ vary between about 1 and 100, and have an average value of 50 g cm^{-1} sec^{-1} in the free atmosphere.

For laminar flow conditions with a vanishingly small pressure difference, Prandtl states the value $\nu/a = 1$. The quantity $\nu = \mu/\rho$ is 0.133 cm^2sec^{-1} at 0°C, and 0.245 cm^2sec^{-1} at 100°C. On the other hand, the quantity $a = \lambda/(\rho \cdot c_p)$ is 0.173 to 0.183 cm^2sec^{-1} if it is assumed that λ (coefficient of thermal conductivity of air) is 0.0000568 cal $cm^{-1}degree^{-1}sec^{-1}$, ρ (density or air) is 0.001293 g cm^{-3}, and c_p (specific heat at constant pressure) is 0.2405 cal $g^{-1}degree^{-1}$. Using these values for ν and a, we obtain average values of 0.7 to 0.8 for ν/a. Pohlhausen used a value of $\nu/a = 0.733$ for air (see Prandtl [33, p. 366].

These considerations and assumptions yield the firm numerical values listed in Table 3.

These values of the constants allow us to estimate the average properties of Prandtl's solutions, which are expressions (8), (9), and (10).

First, it is of interest to determine the altitude n at which the velocity has a maximum. The condition for maximum velocity is $\partial w/\partial n = 0$.

We have

$$\frac{\partial w}{\partial n} = C \cdot \sqrt{\frac{g\beta a}{B\nu}} \cdot e^{-\frac{n}{\ell}} \cdot \left[\cos\frac{n}{\ell} - \sin\frac{n}{\ell}\right] = 0 \quad (11)$$

which gives

$$n = \frac{\pi}{4} \cdot \ell$$

for the altitude at which the velocity is maximum.

However, this altitude is known from the curves shown above in Figures 9 and 10; it is equal to 27 m for both curves (upslope wind and downslope wind).

Thus, using equation (12), the quantity ℓ is calculated as

$$\ell = 34.37 \text{ m} = 3437 \text{ cm} . \quad (13)$$

Furthermore, it is of interest to determine the height at which w = 0. This is given by

Table 3.

Constant	Numerical Value	Units	
α	42.5°	degrees	Turbulent state
g	980.6	$cm^2\ sec^{-1}$	$\frac{a}{\nu} = \frac{A_q}{A_\tau} = 1.4 \sim 1.5$
B	0.4°/100 m	$degree \cdot m^{-1}$	
	4×10^{-5}	$degree \cdot cm^{-1}$	
β	1/283°	degrees	Laminar state
ρ (0m)	1.225×10^{-3}	$g\ cm^{-3}$	$\frac{a}{\nu} = 0.733$
ρ (20000m)	0.09540×10^{-3}	$g\ cm^{-3}$	a $0.183 \sim 0.173\ cm^2\ sec^{-1}$
			ν $0.245\ cm^2\ sec^{-1}$ at 100°C
			$0.133\ cm^2\ sec^{-1}$ at 0°C

$$n_0 = \pi \cdot \ell \qquad (14)$$

and using ℓ = 34.37 m we have

$$n_0 = 108 \text{ m}.$$

Using equation (9), which is

$$\ell = \sqrt[4]{\frac{4\ a\nu}{g\beta B \sin^2\alpha}}$$

and $a/\nu \simeq 1.4$, it is now possible to calculate A_q and A_τ, and we obtain, using the constants in Table 3,

$$A_q = 52.9 \text{ and}$$
$$A_\tau = 37.8\ (g\ cm^{-1} sec^{-1}). \qquad (15)$$

The order of magnitude of these values is well within the range of possibility. The values apply for a slope inclination of 42.5°, which is the average slope inclination of the Nordkette at the altitude of the Seegrube. For other sites, where the average inclination was only 18°, the downslope wind measurements yielded the following values:

$$A_q = 24.2 \text{ and}$$
$$A_\tau = 16.9\ (g\ cm^{-1} sec^{-1}). \qquad (15a)$$

We now turn our attention to numerical evaluation of equation (10), which gives the vertical distribution of velocity w above the slope. At an altitude of 27 m, at which the maximum of w occurs for the upslope wind and for the downslope wind, w is 3.9 m/sec for the upslope wind and -2.36 m/sec for the downslope wind.

The value of $e^{\frac{-n}{\ell}} \cdot \sin \frac{n}{\ell}$ is 0.33 for n = 27 m and ℓ = 34.37 m, and we obtain the values

$$C \cdot \sqrt{\frac{g\beta a}{B\nu}} = 1182 \text{ cm/sec for upslope wind, and}$$
$$= -715 \text{ cm/sec for downslope wind.}$$

The value of $\sqrt{g\beta a/B\nu}$ can be determined, and thus C can be calculated. It is

$$C = 3.4^0 \text{ for upslope wind, and}$$
$$C = -2.05^0 \text{ for downslope wind.}$$

Now it is possible to evaluate the whole expression in equation (10) for various altitudes n (every 10 m from n = 10 m to n = 130 m).

Thus we obtain the theoretical vertical distribution of w in the case of the upslope wind and the case of the downslope wind; the numbers are given in Table 4.

We now return to Figures 9 and 10 and consider the curves for w obtained by plotting the theoretical values (from Prandtl's theory) given in Table 4. If we compare these theoretical curves with the curves of the mean distributions obtained from observations, we see that the theoretical curves are in surprisingly good agreement with the mean curves obtained from observed data, up to an altitude of 27 m (maximum velocity w). The differences in velocity which occur in the altitude range 0 to 27 m are at most 0.4 m/sec, which is within the experimental error of the method of measurement used, and is thus unimportant. As expected, however, above the altitude n = 27 m the theoretical curve does not follow the observed curves, but decreases up to an altitude of n_0 = 108 m, where w = 0, and then even attains moderate negative values. Because of the disturbing influence of the gradient wind (or the mountain and valley winds) mentioned above, the mean curves of the observations are not in complete agreement with the theoretical curves. In order to make the theoretical and observed curves agree, we must subtract a velocity distribution at the individual altitudes (represented by the broken line in the two figures) from the curve of the observed values. In both cases a small deviation becomes noticeable above an altitude of 30 m, but the magnitude of this deviation from the pure slope wind does not become considerable until the altitude reaches 70 to 80 m, and higher. However, the influence of the overlying gradient wind (or even the mountain and valley wind) is very plausible at altitudes about 100 m above that of the Seegrube observation station (1905 m).

In general, the comparison of theory and observation is very satisfactory within the layer of pure slope wind.

We now turn to the task of evaluating equation (8) (vertical temperature distribution). In this case, because of the lack

Table 4.

n Meters	$e^{\frac{-n}{\ell}} \cdot \sin \frac{n}{\ell}$	w (m/sec) Upslope Wind	w (m/sec) Downslope Wind
10	0.1703	+2.01	-1.21
20	0.3070	+3.62	-2.20
27	0.3300	+3.90	-2.36
30	0.3276	+3.87	-2.34
40	0.2872	+3.39	-2.05
50	0.2321	+2.74	-1.66
60	0.1722	+2.03	-1.23
70	0.1169	+1.38	-0.84
80	0.0711	+0.84	-0.51
90	0.0366	+0.43	-0.26
100	0.0127	+0.15	-0.09
108	0.0000	0.0	0.0
110	-0.0024	-0.03	+0.02
120	-0.0104	-0.12	+0.08
130	-0.0136	-0.13	+0.10

of suitable data, we must be satisfied with the theoretical picture of the distribution of potential temperature as a function of altitude n (normal to the slope).

The calculated values of θ' for various altitudes n (every 10 m from n = 10 m to n = 130 m) are given in Table 5 for the upslope wind and the downslope wind.

If the values of θ' in Table 5 are plotted as a function of altitude (as in Figure 12), it can be seen that the value of θ' decreases rapidly in the lower layers and becomes zero at a certain altitude.

The altitude at which $\theta' = 0$ is $n = \pi\ell/2 = 54.05$ m. The layer of temperature anomaly θ' is therefore only half as high as the layer of slope wind (w).

At altitudes higher than the altitude for which $\theta' = 0$, the curve for the upslope wind becomes negative and the curve for the downslope wind becomes positive. With increasing altitude, however, the curves change very little.

Table 5.

Altitude n Meters	Potential Temperature Anomaly θ' °C Upslope Wind	°C Downslope Wind
0	3.4	-2.05
10	2.43	-1.47
20	1.59	-0.96
30	0.91	0.55
40	0.42	-0.25
50	0.09	-0.03
60	-0.10	0.05
70	-0.20	0.12
80	0.23	0.14
90	-0.21	0.13
100	-0.18	0.11
108	-0.15	0.09
110	-0.14	-0.08
120	-0.12	0.07
130	-0.06	0.04

This means that, in the case of the upslope wind and in the case of the downslope wind, the temperature anomaly of the temperature layer above the slope theoretically must extend up to altitudes of about 50 m.

The correct temperature distribution above the slope is obtained from the theory by superimposing the temperature anomaly on the mean temperature distribution, i.e. by evaluating equation (1) using the given numerical values.

This is done in Figures 13 and 14 for the upslope and downslope winds, respectively.

It can be seen from Figure 13 (upslope wind) that in the lower layers the potential temperature changes rapidly with altitude; in fact, there are superadiabatic gradients, since the potential temperature increases [*sic*, decreases] with increasing altitude, up to about 70-80 m (measured normal to the slope). At higher altitudes the temperature distribution gradually approaches the normal decrease in temperature with altitude of 0.6°/100 m.

If we follow the course of an individual isotherm, which is flat at some distance from the slope, we see that the isotherm begins to rise somewhat in a certain region in the neighborhood of the slope; then it merges with the upslope arches caused by the slope wind. This phenomenon was mentioned above.

Figure 14 (downslope wind) has a completely different form. A ground inversion is dominant up to an altitude of about 70-80 m (normal to the slope), beyond which the temperature distribution is rather rapidly converted into the normal decrease of 0.6 degrees/100 m. This theoretical temperature distribution is qualitatively confirmed by the observations.

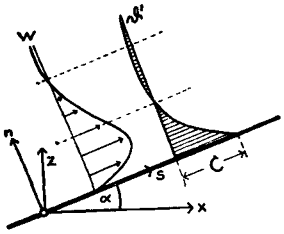

Figure 11: Prandtl's schematic representation of the wind and temperature profiles (normal to the slope).

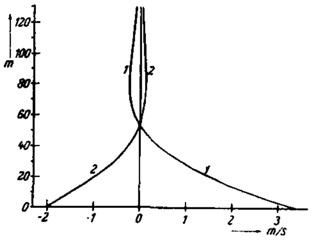

Figure 12: Distribution (normal to the slope) of the temperature anomaly θ' for the upslope wind (1) and the downslope wind (2).

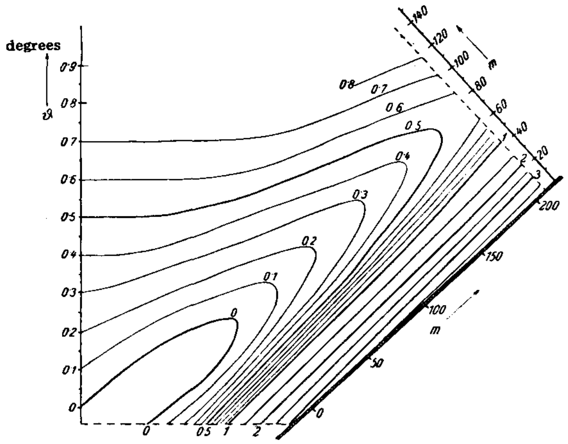

Figure 13: Distribution (normal to the slope) of the potential temperature above the slope (upslope wind).

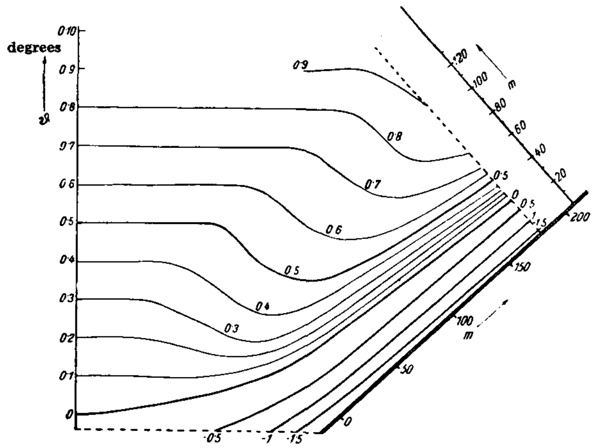

Figure 14. Distribution (normal to the slope) of the potential temperature above the slope (downslope wind).

In this comparison of theory with observations, it should be noted that the theoretical calculations of the wind and temperature distributions always used coefficients of friction and heat exchange which were constant throughout the entire layer. In reality, this assumption represents natural conditions only in a rough way. It may be that variation of these coefficients with altitude has less effect on the external form of the wind and temperature distributions (normal to the slope) than on the thickness of the layer in which the disturbances of wind and temperature are effective. There is no doubt that, to a first approximation, Prandtl's theory gives useful insight into the whole mechanism of wind and temperature disturbances involved in slope winds.

V. <u>Expanded theory of periodic diurnal slope wind currents; discussion of expanded theory</u>

We now consider the question of whether the nonstationary case can also be treated by a suitably modified form of Prandtl's theory. In the nonstationary case, $\partial w/\partial t$ and $\partial \Theta/\partial t$ are no longer to be set equal to zero, and we obtain the equations

$$\frac{\partial w}{\partial t} = g\beta \cdot \sin\alpha \cdot \Theta + \nu \frac{\partial^2 w}{\partial n^2}$$

and

$$\frac{\partial \Theta}{\partial t} = - B \cdot \sin\alpha \cdot w + a \frac{\partial^2 \Theta}{\partial n^2} \quad . \tag{17}$$

which correspond to equations (2) and (6). If we define $g\beta \cdot \sin\alpha = A_1$ and $B \sin\alpha = A_2$, then

$$\frac{\partial w}{\partial t} - A_1 \Theta - \nu \frac{\partial^2 w}{\partial n^2} = 0,$$
$$\frac{\partial \Theta}{\partial t} + A_2 w - a \frac{\partial^2 \Theta}{\partial n^2} = 0. \tag{18}$$

Θ and w no longer depend only on n, but also depend on the time t with a period σ = 24 hours (i.e. the diurnal period, since the anomaly Θ exhibits such a diurnal period).

Therefore it is appropriate to set

$$\Theta = Ee^{(i-1)\cdot kn + i\sigma t},$$
$$w = -i\, De^{(i-1)\cdot kn + i\sigma t}. \tag{19}$$

so that

$$\Theta = -\frac{E}{iD} \cdot w \; ; \; w = -\frac{iD}{E} \cdot \Theta \quad .$$

D and E are arbitrary constants.

Inserting these expressions in the differential equations (18) gives

$$i\sigma + \frac{A_1 E}{iD} + 2\, ik^2 \cdot \nu = 0,$$
$$i\sigma - \frac{iA_2 D}{E} + 2\, ik^2 \cdot a = 0, \tag{20}$$

and, using $E/D = x$, we obtain the following pair of equations:

$$\sigma - A_1 \cdot x + 2\, k^2 \cdot \nu = 0,$$
$$\sigma - \frac{A_2}{x} + 2\, k^2 \cdot a = 0. \tag{21}$$

These two equations are sufficient to determine k and x. Before doing this, however, we shall clarify a few intermediate questions.

a) For the case $\sigma = 0$ we must obtain the stationary-state Prandtl solutions for w and Θ. Inserting the condition $\sigma = 0$ in equations (21), we obtain

$$\frac{A_1}{A_2} \cdot x^2 = \frac{\nu}{a} \text{ and thus } x = \sqrt{\frac{B\nu}{ag\beta}} = \frac{E}{D};$$

$$k^2 = \frac{A_1 \cdot x}{2\nu} \text{ and thus } k = \sqrt[4]{\frac{g\beta B \sin^2\alpha}{4\nu a}} = \frac{1}{\ell} \, .$$

From equations (19), using $\sigma = 0$, $D \cdot \sqrt{B\nu/ag\beta} = E$ = the Prandtl constant C, and $k = 1/\ell$, we obtain

$$\Theta = Ce^{\frac{in}{\ell}} \cdot e^{\frac{-n}{\ell}} = Ce^{\frac{-n}{\ell}} \cdot \cos\frac{n}{\ell} \; ,$$

$$w = -\; iC \cdot \sqrt{\frac{ag\beta}{B\nu}} \cdot e^{\frac{in}{\ell}} \cdot e^{\frac{-n}{\ell}}$$
$$= C \cdot \sqrt{\frac{ag\beta}{B\nu}} \cdot e^{\frac{-n}{\ell}} \cdot \sin\frac{n}{\ell} \, .$$

The real parts give Prandtl's w and Θ, as expected.

b) It is interesting to see how much the solution deviates from the Prandtl solution if we assume that the quantity $\nu/a = 1$, which, as we know, is not strictly the case (see [33, p. 366].

If the condition ν = a is inserted in equations (21), we obtain

$$x = \sqrt{\frac{A_2}{A_1}} = \sqrt{\frac{B}{g\beta}} \, , \tag{23}$$

which is identical to the expression for the case $\sigma = 0$ with $\nu = a$.

On the other hand,

$$k^2 = \frac{\sqrt{A_1A_2} - \sigma}{2\nu} = \sqrt{\frac{g\beta b \sin^2\alpha}{4\nu^2}} - \frac{\sigma}{2\nu} = \frac{1}{\ell^2} - \frac{\sigma}{2\nu}$$

or

$$k = \frac{1}{\ell} \cdot \sqrt{1 - \frac{\sigma\ell^2}{2\nu}} \quad . \qquad (24)$$

This expression for k is different from the value of k for $\sigma = 0$ because of the term $\sigma\ell^2/2\nu$ under the radical sign.

Using the values $\sigma = 2\pi/24$ hours $= 2\pi/86{,}400$ sec, $\ell = 34.37$ m, $\nu = A_\tau/\rho = 37.77/1.293 \times 10^{-3}$, the value of the term $\sigma\ell^2/2\nu$ turns out to be 0.014704, and thus

$$k = \frac{1}{\ell} \cdot \sqrt{1 - 0.014704}\,, \qquad (25)$$

which is approximately equal to $1/\ell$; i.e., the value of k for $\nu = a$ is approximately equal to the value of k for $\sigma = 0$.

Thus we have established that the choice of $\nu = a$ (i.e. $\nu/a = 1$) as opposed to $\nu/a \neq 1$ (i.e. $\nu/a \simeq 0.7$) does not play an important role for the solutions. This result is of some interest.

c) We now consider the values of x and k for the case $\nu \neq a$ and compare them with the values of x and k for $\sigma = 0$.

With $\nu \neq a$, equation (21) gives the following quadratic equation for determination of x:

$$x^2 - \frac{\sigma}{A_1}\left(1 - \frac{\nu}{a}\right) x - \frac{A_2}{A_1} \cdot \frac{\nu}{a} = 0$$

or

$$x = + \frac{\sigma}{2A_1}\left(1 - \frac{\nu}{a}\right) \pm \sqrt{\frac{\sigma^2}{4A_1^2}\left(1 - \frac{\nu}{a}\right)^2 + \frac{A_2}{A_1} \cdot \frac{\nu}{a}} \quad . \qquad (26)$$

Calculation of the value of $(\sigma/2A_1) \cdot (1 - \nu/a)$ yields 4.44×10^{-6}, which is an extraordinarily small number. Therefore the quadratic term under the radical sign is completely negligible in comparison with the term $A_2/A_1 \cdot \nu/a$, and we have

$$\sqrt{\frac{A_2}{A_1} \cdot \frac{\nu}{a}} = 2.871 \times 10^{-3} \quad .$$

Therefore this term is by far the main contribution to the value of x, and we can (to a very close approximation) set

$$x = \sqrt{\frac{A_2}{A_1} \cdot \frac{\nu}{a}} = \sqrt{\frac{B\nu}{g\beta a}} \quad . \qquad (27)$$

Thus for the case $\nu \neq a$ there is no essential difference in the value of x in comparison to the value of x in the stationary case ($\sigma = 0$) of Prandtl.

For the determination of k when $\nu \neq a$, we obtain from equation (21) the following equation:

$$k^2 = \frac{\sqrt{\frac{A_2}{A_1} \cdot \frac{\nu}{\alpha}}}{2\nu} - \frac{\sigma}{2\nu} = \sqrt{\frac{g\beta B \sin^2\alpha}{4\,\nu a}} - \frac{\sigma}{2\nu}$$
$$= \frac{1}{\ell^2} - \frac{\sigma}{2\nu} \qquad (28)$$

or

$$k = \frac{1}{\ell} \cdot \sqrt{1 - \frac{\sigma\ell^2}{2\nu}} \approx \frac{1}{\ell} \;.$$

We have already shown that this value for k is, to a very close approximation, equal to $1/\ell$, and thus in the case $\nu \neq a$ the value of k is equal to the value of k for the case $\sigma = 0$. In doing this, however, we have proved that, to a very close approximation, our solutions for the potential temperature anomaly θ and the velocity w are, <u>under any assumptions for ν and a</u>

with regard to $k = 1/\ell$ and $x = E/D = \sqrt{B\nu/ag\beta}$, exactly the same in the case $\sigma \neq 0$ (periodic diurnal variation of slope winds) as they are in Prandtl's stationary case $\sigma = 0$.

Therefore the complete solutions are

$$\Theta = C \cdot e^{\frac{in}{\ell}} \cdot e^{\frac{-n}{\ell}} \cdot e^{i\sigma t} , \qquad (29)$$

$$w = - iC \cdot \sqrt{\frac{ag\beta}{\nu B}} \cdot e^{\frac{in}{\ell}} \cdot e^{\frac{n}{\ell}} \cdot e^{i\sigma t} .$$

Considering only the real parts, we have

$$\Theta = C \cdot e^{\frac{-n}{\ell}} \cdot \cos \frac{n}{\ell} \cdot \cos \sigma t = \underline{\Theta_{stat} \cdot \cos \sigma t},$$

$$w = C \cdot \sqrt{\frac{ag\beta}{\nu B}} \cdot e^{\frac{-n}{\ell}} \cdot \sin \frac{n}{\ell} \cdot \cos \sigma t \qquad (30)$$

$$= \underline{w_{stat} \cdot \cos \sigma t.}$$

Equations (30) show that, if Prandtl's slope wind theory is applied to the nonstationary case (assuming a periodic diurnal slope wind circulation), then the Prandtl solutions for the potential temperature anomaly θ and the velocity w (parallel to the slope) remain substantially unchanged: the solutions for θ and w are merely to be multiplied by the periodic factor $\cos \sigma$. Even if ν and a are changed in value, the constants E, D, x, and k can always be replaced in the calculation by the Prandtl constants C and ℓ, to a close approximation.

This simple result, which is obtained under the assumption of periodic slope winds (which is essentially more complicated than the assumption of a stationary state) is certainly of some interest, and is noteworthy.

VI. Comparison of Prandtl's result with A. Defant's work concerning the flow of cold air masses downslopes

I shall now discuss the work of my father concerning the flow of heavy air masses downslopes, along with his theoretical considerations concerning the stability conditions for stationary air flows of this type. Let a two-layer atmosphere of finite height H be constructed above a slope. Suppose that the potential temperature θ of the lower layer is constant, and that the potential temperature θ_1 of the upper layer is also constant. Define n to be the altitude of the boundary surface between these two differently constructed layers.

Assume that the altitude H of the upper boundary of the upper layer is so large in comparison to the height n of the lower layer that the upper boundary can always be thought of as being horizontal; then we have

$$- \frac{1}{\rho} \frac{\partial p}{\partial x} = - g \cos \alpha \frac{\Theta_1 - \Theta}{\Theta_1} \cdot \frac{\partial n}{\partial x} - g \frac{\Theta}{\Theta_1} \cdot \sin\alpha, \qquad (31)$$

where x is the coordinate (lying in the plane of the slope) which is positive in the downward direction (x is identical to the coordinate -s, used above).

Using equation (31) and the equations of motion which apply for an appropriate coordinate system (x parallel to the slope, z perpendicular to the slope), we obtain the following equation for the gradient force in the x-direction:

$$\frac{dw}{dt} = g \cos \alpha \frac{\Theta_1 - \Theta}{\Theta_1} \left[tg\, \alpha - \frac{\partial n}{\partial x} \right]. \qquad (32)$$

This equation is of the greatest interest, since it permits us to estimate (by comparing the terms inside the brackets) the effect of the inclination of the surface (tg α) and the effect of the change (in the x-direction) of the altitude η of the boundary surface ($\partial\eta/\partial x$). If the boundary surface is parallel to the slope, then only the gravity term is effective, since $\partial\eta/\partial x = 0$. If $\partial\eta/\partial x$ is positive or negative, then the effect of the gravity term is reduced or increased, respectively. Comparison of the order of magnitude of tg α with that of $\partial\eta/\partial x$ yields the result that the two terms can be of equal magnitude, which means that it is not permissible to neglect one in comparison to the other. If we assume that the frictional force is given by the Taylor formula (for the whole column of air above a unit surface area of the ground) in the form $\mu \cdot w^2$ (μ = 0.002 for ground which is not too rough), then for a stationary air flow along the slope we obtain (using the assumptions $\eta = Z$ = constant, $w = W$ = constant) an equation for the velocity W parallel to the slope:

$$W = \left(\frac{gZ}{\mu} \cdot \frac{\Theta_1 - \Theta}{\Theta_1} \cdot \sin \alpha\right)^{\frac{1}{2}} . \quad (33)$$

The difference between this (A. Defant's) result and Prandtl's result is that Prandtl's equation (for the stationary air flow parallel to the slope) permits a vertical velocity distribution to be obtained, whereas the above expression (33) merely gives an average measure of the velocity along the slope in the whole lower layer.

The theoretical foundations of the two works are indeed the same, but Prandtl uses a stratified atmosphere whose stratification on the slope is perturbed by the transfer of heat from the warm surface of the slope, whereas A. Defant works with a two-layer atmosphere in which the assumed difference in the potential temperatures of the two layers is the cause of the motion of the lower layer. However, the presence of the boundary surface between the lower and upper layers allows us to consider a pressure drop along this slope. This possibility does not exist in Prandtl's model. Thus it is possible (in A. Defant's theory) to estimate the relative effects of this pressure drop (acting from above) and of gravity on the motion of the air in the lower layer.

In connection with this, there is the further possibility of investigating in detail the stability of an air flow of this type, and A. Defant determined (in agreement with H. Jeffreys) that the downflowing air current becomes unstable if the inclination of the slope becomes greater than four times the coefficient of friction (tg $\alpha \geq 4\mu$).

It is not easy to compare the velocities given by the two theories because of the great differences in the foundations of the two theories. In particular, the work of A. Defant uses the Taylor friction formula (shear stress $\tau = \mu \cdot \rho \cdot w^2$) for the entire lower layer, where the coefficient of friction μ can assume quite different values depending on the nature of the surface. If we wish to use expression (33) (for W) in calculating the stationary velocity parallel to the slope, then we must make assumptions about the following three things: the layer thickness Z, the temperature difference $\Theta_1 - \Theta$ between the two layers, and the value of the coefficient of friction. We can also proceed in the following manner: We can select an average value of the velocity profile of the Prandtl solution (see Figures 9 and 10) to correspond to the average velocity of the entire lower layer, as well as selecting an average value of the temperature anomaly at the same altitude; then we can calculate the value of μ using equation

(33). I have performed exact calculations of this type, and they all lead to values of μ of the order of 0.05 to 0.10. This indicates that the turbulence of this flow must be rather great, which is understandable if we consider that a decidedly unstable state is present in the structure of the two layers when the potential temperature Θ_1 of the upper layer is greater than the potential temperature of the lower layer. Furthermore, we must remember that the slopes on which the observations were carried out were rather rough: on the one hand, they were overgrown with alpine forest, and on the other hand the slope surface was not uniform because of ditches and furrows.

Moreover, for the turbulent case, Prandtl himself has shown (see [33, p. 375]) that the assumption that shear stress τ at the rough slope is of the form $\tau = \mu\rho w^2{}_1$ leads to the same values of velocity as does the previous assumption in the Prandtl solution.

The thermal buoyancy of the entire heated layer of thickness ℓ must be set approximately equal to the shear stress τ. This gives

$$\mu\rho w_1^2 \sim g\,\rho\,\sin\alpha \cdot\beta\cdot C\cdot\ell \qquad (34)$$

where w_1 = maximum velocity.

Now $\tau = A_\tau \cdot \partial w/\partial n \sim A_\tau\, w_1/\ell = \mu\rho w_1^2$, so that $A_\tau \sim \rho\cdot w_1\cdot\ell\cdot\mu$.

Further, from equation (6), if we set $a = A_q/\rho$, we obtain

$$w_1 \cdot B \sin a \sim \frac{A_q}{\rho} \cdot \frac{C}{\ell^2} \quad .$$

The two preceding relations lead to the relation

$$\ell \sim \frac{A_q}{A_\tau} \cdot \frac{C}{B\sin\alpha} \cdot \mu \qquad (35)$$

This expression, along with equation (34), then leads to

$$w_1 \sim C \cdot \sqrt{\frac{A_q}{A_\tau} \cdot \frac{g\beta}{B}} \quad . \qquad (36)$$

However, the content of this is identical to equation (10) if we replace the term a/ν in equation (10) by the ratio of the exchange values, A_q/A_τ.*

Thus the turbulent case does not lead to velocities which are different from those of the laminar case.

VII. Remarks on the theory of mountain and valley winds

In connection with his theory of slope winds, Prandtl remarks that the currents on mountain slopes (i.e. slope winds), which flow up the slope in the case of heating and down the slope in the case of cooling, create (at the valley floor and at the ridges) "sinks" and "sources" for the free current. These sinks and sources must have an effect on the free currents along the valley. Thus, for example, if a long valley between high mountains is filled with stably stratified air, then the air which is removed by the sinks at the foot of the heated slopes can be replaced only by air that flows into the valley from the valley exit. This air can move only in the horizontal direction because of the stratification.

Prandtl's explanation of mountain and valley winds can be valid only if the air masses ascending along the slopes are

---

* In Prandtl's derivation the coefficient of friction μ (ζ in Prandtl's notation) is under the radical sign instead of A_q/A_τ; this is attributable to an error in calculation.

completely (or mostly) transported away and do not return to the valley. According to the observations and theory of A. Wagner, this is not the case at all times of day. On the contrary, a closed slope wind circulation (see Figure 2) is unconditionally required to generate the plain-valley pressure drop at the time of heating. Only in this way can the cold air column be replaced or expelled by heated slope wind air during the period from morning until noon. Thus, with the same pressure at the altitude of the ridge, there is a colder air mass over the plain next to a warmer air mass over the valley floor. This results in a plain-valley pressure drop. The assumption of sinks and sources is suitable only for a part of this phenomenon. The situation is somewhat different in the case of the mountain wind, because the air flowing down over the valley from the slopes can not ascend easily, and will flow along the valley floor and off toward the plain.

The observed and calculated slope wind profiles make it possible to estimate, for the case of closed slope wind circulation, the period of time in which the ascending slope air is capable of replacing the air masses above the valley floor. According to equation (8), the heating effect on the slope takes place up to an altitude of $n = (\pi/2) \cdot \ell$. If we use equation (10) (for w) to calculate the mean upslope wind velocity of this layer, then simple integration from the slope surface up to the altitude $(\pi/2) \cdot \ell$ gives us the relation

$$W_m = \frac{C}{\pi} \cdot \sqrt{\frac{g\beta a}{B\nu}} \cdot \left[1 - e^{\frac{-\pi}{2}} \right] . \qquad (37)$$

If we choose $C = 1182$ cm/sec (see p. 109), then we obtain $W_m = 2.98$ m/sec for the mean upslope wind velocity.

For a layer thickness of $(\pi/2) \cdot \ell = 54.05$ m (see p. 111) and a width of 1 m, we obtain 161 m^3/sec (i.e. 5.8×10^5 m^3/hour) as the transport rate through this cross-section. This is the rate at which air heated on the slope enters (at the altitude of the crest and below) the column of air above the valley floor. If we assume that this column of air is about 1750 m high, which corresponds to the Inn valley, and has a base area of 1500 m x 1 m = 1500 m^2 (1500 m corresponds to half the width of the valley floor), then we obtain a volume of about 2.6×10^6 m^3.

If this volume above half the valley floor is divided by the above-mentioned transport rate, we obtain

$$\frac{\text{volume}}{\text{transport rate}} = \frac{2.6 \times 10^6}{5.8 \times 10^5} \frac{m^3}{m^3/\text{hour}} \qquad (38)$$

$$= 4.5 \text{ hours} .$$

This means that the air mass above the valley floor is replaced in about 4 1/2 hours, assuming closed circulation of the warm slope air.

If we assume that the heating effect begins at about 8:30 AM, then the greatest pressure difference between plain and valley would occur at about 1 PM; at this time the main phase of development of the up-valley wind would begin. Since air masses extending from the valley exit to Innsbruck, for example, must first be set in motion, the greatest development of the up-valley wind at Innsbruck will take place between 3 PM and 4 PM. This is in agreement with observations (see [4, Figure 1]). This crude estimate is applicable only under the assumption that the slope wind circulation is closed. If a portion of the air of the upslope wind is removed at the upper parts of the valley, then the process takes a correspondingly longer time. Nevertheless, this calculation shows that the return of slope wind air to the middle of the valley is highly significant for the theory of mountain and valley winds.

A really complete theory of slope winds would also take into consideration the descending branch in the middle of the valley. The continuity equation would have to be used in addition to the other equations in order to treat the closed circulation, and we would arrive at a problem similar to the one encountered in the case of heat convection over a flat plate (see [33, p. 371]). The flat plate, however, would have to be replaced by adjacent inclined surfaces corresponding to mountain and valley; these surfaces would determine the size of the convection cell. However, mathematical difficulties might well be encountered in the theoretical treatment of such a system.

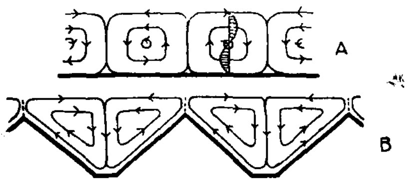

Figure 15: A: Thermal convection cells above a heated horizontal plate. B: Slope wind convection cells above a mountainous region (schematic).

References

1. H. Tollner: Berg- und Talwinde in Osterreich [Mountain and valley winds in Austria]. Jb. Zentr.-Anst. f. Meteorol. u. Geod. Wien, Beiheft zu Jg. 1928, 91.

2. R. Wenger: Zur Theorie der Berg- und Talwinde [A theory of mountain and valley winds]. Meteorol. Z. 1923, 193.

3. H. Tollner: Gletscherwinde in den Ostalpen [Glacier breezes in the East Alps]. Meteorol. Z. 1931, 414.

4. E. Ekhart: Zur Aerologie des Berg- und Talwindes [The aerology of mountain and valley winds]. Beitr. Physik. fr. Atm. 18, 1 (1932).

5. E. Ekhart: Weitere Beitraege zum Problem des Berg- und Talwindes [Further contributions to the problem of mountain and valley winds]. Beitr. Physik. fr. Atm. 18, 242 (1932).

6. E. van Everdingen: Zur Theorie der Berg- und Talwinde [A theory of mountain and valley winds]. Beitr. Physik fr. Atm. 19, 109 (1932).

7. A. Wagner: Hangwind -- Ausgleichsstroemung -- Berg- und Talwinde [Slope winds; equalization currents; mountain and valley winds]. Meteorol. Z. 1932, 209.

8. A. Wagner: Der taegliche Luftdruck und Temperaturgang in der freien Atmosphaere und in Gebirgstaelern [The diurnal course of temperature and air pressure in the free atmosphere and in valleys]. Gerlands Beitr. Geophysik, 37, 315 (1932).

9. A. Wagner: Neue Theorie des Berg- und Talwindes [New theory of mountain and valley winds]. Meteorol. Z. 1932, 331.

10. C. Braak: Der Malojawind [The Maloja wind]. Meteorol. Z. 1933, 231.

11. A. Roschkott, F. Steinhauser u. F. Lauscher: Winduntersuchungen im Sonnblickgebiet [Wind studies in the Sonnblick area]. 42. Jb. Sonnblickverein 1933, 15.

12. A. Jelinek: Untersuchung periodischer Tageswinde in Suedtirol [Study of periodic diurnal winds in the South Tyrol]. Beitr. Physik fr. Atm. 21, 223 (1934).

13. E. Ekhart: Neuere Untersuchungen zur Aerologie der Talwinde: Die periodischen Tageswinde in einem Quertale der Alpen [New studies of the aerology of valley winds: the periodic diurnal winds in a transverse valley of the Alps]. Beitr. Physik fr. Atm. 21, 245 (1934).

14. F. Lauscher und O. Eckel: Pilotbeobachtungen im Gailtal (Kaernten) [Pilot balloon observations in the Gail valley (Kaernten)]. Meteorol. Z. 1934, 201.

15. W. Schwabl: Zur Kenntnis der Beeinflussung der Allgemeinstroemung durch ein Gebirgstal [Factors which affect the general flow of air through a valley]. Meteorol. Z. 1934, 342.

16. E. Ekhart: Über einige Gletscherwindmessungen in den Ötztaler Alpen [Some measurements of the glacier breeze in the Oetztal Alps]. Z. Gletscherkunde 22, 217 (1935).

17. E. Moll: Aerologische Untersuchungen periodischer Gebirgswinde in V-foermigen Alpentaelern [Aerological studies of periodic mountain winds in V-shaped alpine valleys]. Beitr. Physik fr. Atm. 22, 177 (1935).

18. T. Zuchristian: Ergebnisse der Windregistrierungen auf einer Kammstation (Hafelekar 2265 m) [Wind measurements at a summit station. (Hafelekar 2265 m)]. Meteorol. Z. 1935, 177.

19. A. Riedel: Zur Aerologie der periodischen Gebirgswinde an einem Suedhange. Ergebnisse von Doppelvisierungen im Innsbrucker Nordkettengebiet [The aerology of periodic mountain winds on a southern slope. Results of double theodolite sightings in the Innsbruck Nordkette area]. Dissert. Innsbruck (1935).

20. F. Bondy: Über Beziehungen zwischen periodischen Talwinden und verschiedenen meteorologischen Faktoren [Relations between periodic valley winds and various meteorological factors]. Dissert. Innsbruck (1935).

21. E. Schaller: Aerologische Untersuchung des periodischen Talwindsystems von Trient [An aerological study of the periodic valley wind system of Trient]. Dissert. Innsbruck (1936).

22. E. Ekhart: Die Windverhaeltnisse ueber einer inneralpinen Station [Wind conditions above a station in the Alps]. Beitr. Physik fr. Atm. 23, 165 (1936).

23. E. Ekhart: Die Windverhaeltnisse des Arlbergs. I. Teil: Bodenbeobachtungen [Wind conditions of the Arlberg. Part I: Ground observations]. Gerlands Beitr. Geophysik 48, 313 (1936).

24. E. Ekhart: Die Windverhaeltnisse des Arlbergs. II. Teil: Aerologische Beobachtungen [Wind conditions of the Arlberg. Part II: Aerological observations]. Gerlands Beitr. Geophysik 49, 7 (1936).

25. A. Burger und E. Ekhart: Über die taegliche Zirkulation der Atmosphaere im Bereiche der Alpen [Diurnal circulation of the atmosphere in the region of the Alps]. Gerlands Beitr. Geophysik 49, 341 (1937).

26. A. Jelinek: Beitraege zur Mechanik der periodischen Hangwinde [Contributions to the mechanism of periodic slope winds]. Beitr. Physik fr. Atm. 24, 60 (1937).

27. A. Jelinek: Über den thermischen Aufbau der periodischen Hangwinde [The thermal structure of periodic slope winds]. Beitr. Physik fr. Atm. 24, 85 (1937).

28. A. Jelinek und A. Riedel: Über die Schichtdicke der periodischen Lokalwinde im Inntal [The layer thickness of periodic local winds in the Inn valley]. Beitr. Physik fr. Atm. 24, 205 (1937).

29. E. Ekhart: Die taegliche Zirkulation der Atmosphaere im Bereiche der Alpen [Diurnal circulation of the atmosphere in the region of the Alps]. Forsch. u. Fortschr. 1937, 210.

30. E. Ekhart: Die Tageszeitenwinde der Alpen, eine Darstellung nach dem neuesten Stande unseres Wissens. [The diurnal winds of the Alps: A model based on our most recent knowledge]. Naturwiss. 1938, 21.

31. A. Wagner: Theorie und Beobachtung der periodischen Gebirgswinde [Theory and observation of periodic mountain winds]. Gerlands Beitr. Geophysik 52, 408 (1938).

32. A. Defant: Der Abfluß schwerer Luftmassen auf geneigtem Boden nebst einigen Bemerkungen zu der Theorie stationaerer Luftstroeme [The flow of heavy air masses downslopes, along with some remarks on the theory of stationary air currents]. Sitz. Ber. Preuß. Akad. Wiss. Phys.-Math.-Klasse XVIII 1933, 3.

33. L. Prandtl: Stroemungslehre [Flow theory]. Verlag Vieweg & Sohn, Braunschweig 1942.

34. H. Lettau: Atmosphaerische Turbulenz [Atmospheric turbulence]. Akad. Verlagsgesellschaft m. b. H. Leipzig 1939.

35. H. Koschmieder: Dynamische Meteorologie [Dynamic meteorology]. Akad. Verlagsgesellschaft m. b. H. Leipzig 1939.

36. F. M. Exner: Dynamische Meteorologie. 2. Aufl [Dynamic meteorology. Second Edition]. Verlag Julius Springer Wien 1925.

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