Rain and Snow at High Elevation 1

The Interaction of Water, Energy and Trace Substances

Michael Kuhn

1.1 Introduction

Plants are major players in the alpine biogeochemical cycles, using water, energy and nutrients from both the atmosphere and the ground for their primary production. They are exposed to rain and snowfall, may be covered by snow for considerable periods, absorb solar radiation and transpire water vapour back to the atmosphere. While the supply of energy, water and nutrients from the atmosphere is the boundary condition for the plants' existence, they significantly determine the return of all three quantities back to the air.

Bioclimatic temperatures in the high Alps are treated in the chapter by Larcher, the supply of solar radiation by Blumthaler. This chapter deals with the significance of rain and snow for high alpine plants. It describes the regional and local distribution of precipitation, its change with elevation and its seasonal course. It emphasizes the importance of snow as a place of water storage, thermal insulation and concentrated release of ions.

1.2 Regional, Vertical and Seasonal Distribution of Precipitation

1.2.1 Annual Precipitation

The regional distribution of annual precipitation in the Alps has been analysed repeatedly, I recommend reading Fliri (1975), Baumgartner and Reichel (1983), Frei and Schär (1998) and Efthymiadis et al. (2006) for that purpose. All of these authors agree that the distribution of alpine precipitation is dominated by two independent variables: altitude and windward situation (or distance from the northern and southern margins toward the interior Alps).

In the Eastern Alps, the majority of annual precipitation arrives either from the SW or NW, with the passage of a trough on its eastward way from the Atlantic. This explains the frequent succession of south-westerly flow with precipitation at the southern alpine chains followed by north-westerly currents wetting the northern part of the Eastern Alps, a pattern that was described by Nickus et al. (1998): "A trough moving in the Westerlies and moving near the Alps will cause a south-westerly to southerly flow over the Eastern Alps. Precipitation south of the central ridge and Föhn winds in the north are the most frequent weather situation at this stage. With increasing cyclonality air flow will become more westerly, often bringing moist air from the Atlantic. Precipitation will then shift to the central and northern parts of the Alps, starting in the west and continuing to the east. As the flow turns to more northerly directions, a passing cold front may bring precipitation mainly in the northern parts of the Alps and at least an interruption in precipitation in the south."

A consequent rule of thumb is that stations at the northern and southern margins experience three times as much annual precipitation as those in the dry interior valleys (Fliri 1975), in rough figures 1,500–500 mm, with maxima at some stations exceeding 3,000 mm and minima of less than 500 mm per year. In either case, the highest of the central chains experience a secondary maximum as the Glockner Group in Fig. 1.1 (at 47°N and 12.5–13.5°E).

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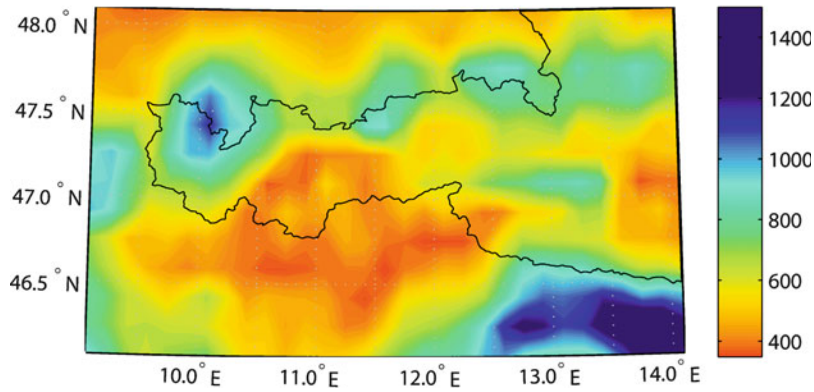

Fig. 1.1 Winter precipitation (December, January, February) in the Eastern Alps according to the HISTALP analysis, in mm. The HISTALP precipitation data set is described by Eftymiadis et al. (2006)

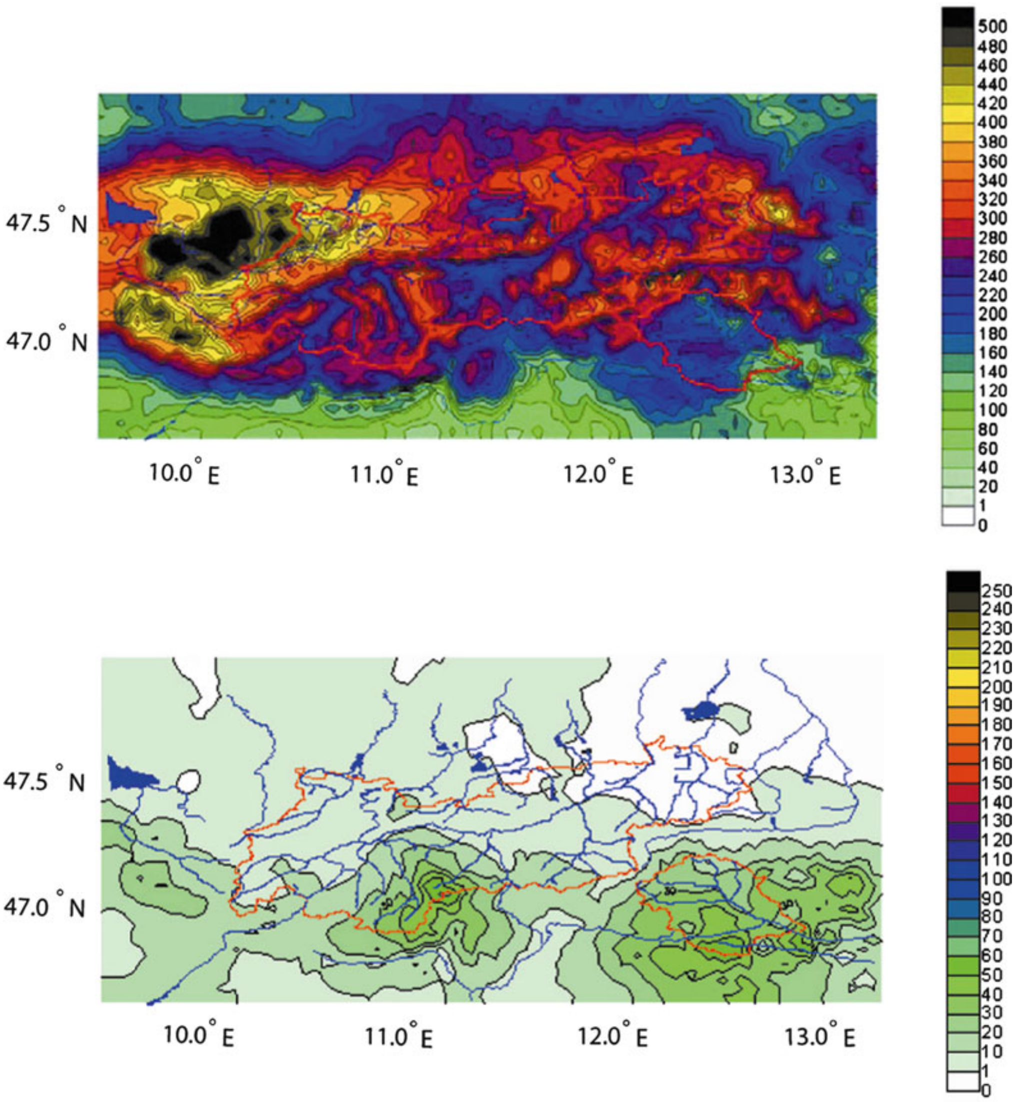

Fig. 1.2 Total precipitation of the month of August 2010 (*above*) is dominated by north-westerly flow with values exceeding 400 mm in the west and less than 200 in the south-east. The daily sums from 14-08-2010, 07:00 to 15-08-2010, 07:00 given in the *lower* panel describe a situation of southerly flow. From Gattermayr (2010)

The annual values given above are the sums of many individual events. Two examples of these are given in Fig. 1.2. Be aware, however, that many of the details on these precipitation maps are interpolated with algorithms that use elevation and distance from actual meteorological observations as independent variables.

1.2.2 Effects of Elevation

Several effects contribute to the higher incidence of precipitation at higher elevations: temperature is lower at higher elevation, the water vapour is thus closer to saturation, and condensation is more likely at higher elevation; when moist air is advected towards a mountain chain, it is forced to ascend and thereby cools; wind speed, and thus horizontal advection of moisture, increases with elevation, the decisive quantity being the product of horizontal wind speed and water vapour density.

Altogether an increase of about 5% per 100 m elevation is observed in various valleys of the Eastern Alps as shown in Fig. 1.3. The annual course of the increase of precipitation with elevation shown in Fig. 1.3 reflects the varying frequencies of advective and convective precipitation. In winter and spring the advective type, associated with the passage of fronts, dominates and leads to high values of the vertical gradient of precipitation. In summer convective precipitation prevails; it depends on local heat sources which are by and large independent of elevation. Convective precipitation most likely occurs over sunlit slopes with vegetation. As both energy supply and vegetation decrease with elevation, convective precipitation may even have an upper limit and thus a lesser increase with elevation; it is certainly more influenced by exposition than by elevation and decreases the mean monthly values of vertical gradients of precipitation given in Fig. 1.3.

1.2.3 Seasonal Variation of Precipitation

The seasonal course of precipitation in the Eastern Alps (e.g. 10°E) has a marked change with latitude. In the north, there is a clear dominance of summer rainfalls, monthly sums may be three times as high as those of fall and winter. A summary presentation by Fliri (1975, his Figs. 69–75) shows this summer maximum to extend southward into the dry, central region. Farther south the Mediterranean dry summers split the precipitation curve into two maxima in spring and fall,

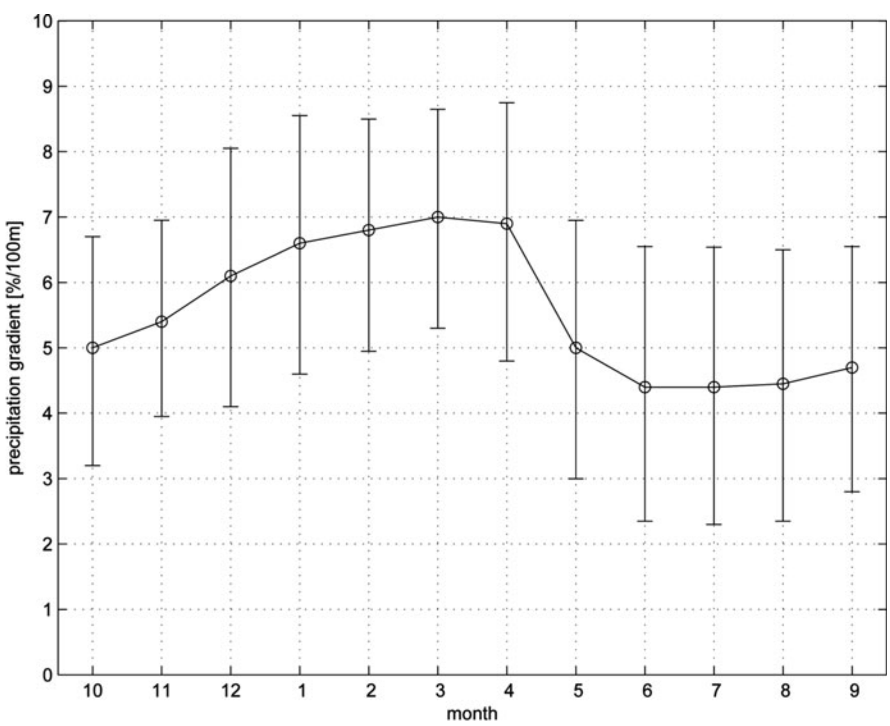

Fig. 1.3 The increase of annual precipitation with elevation, expressed as % per 100 m elevation. Bars indicate upper and lower limits of 10 hydrological basins (Kuhn 2010)

the latter often dominating. This is locally diverse, but generally evident in Fig. 1.4.

An analysis of recent years has shown that the climatological means in Fig. 1.4 have a high interannual variance and that deviations from the mean have a tendency to come in groups of several years. This is true in particular for the occurrence of October and November maxima.

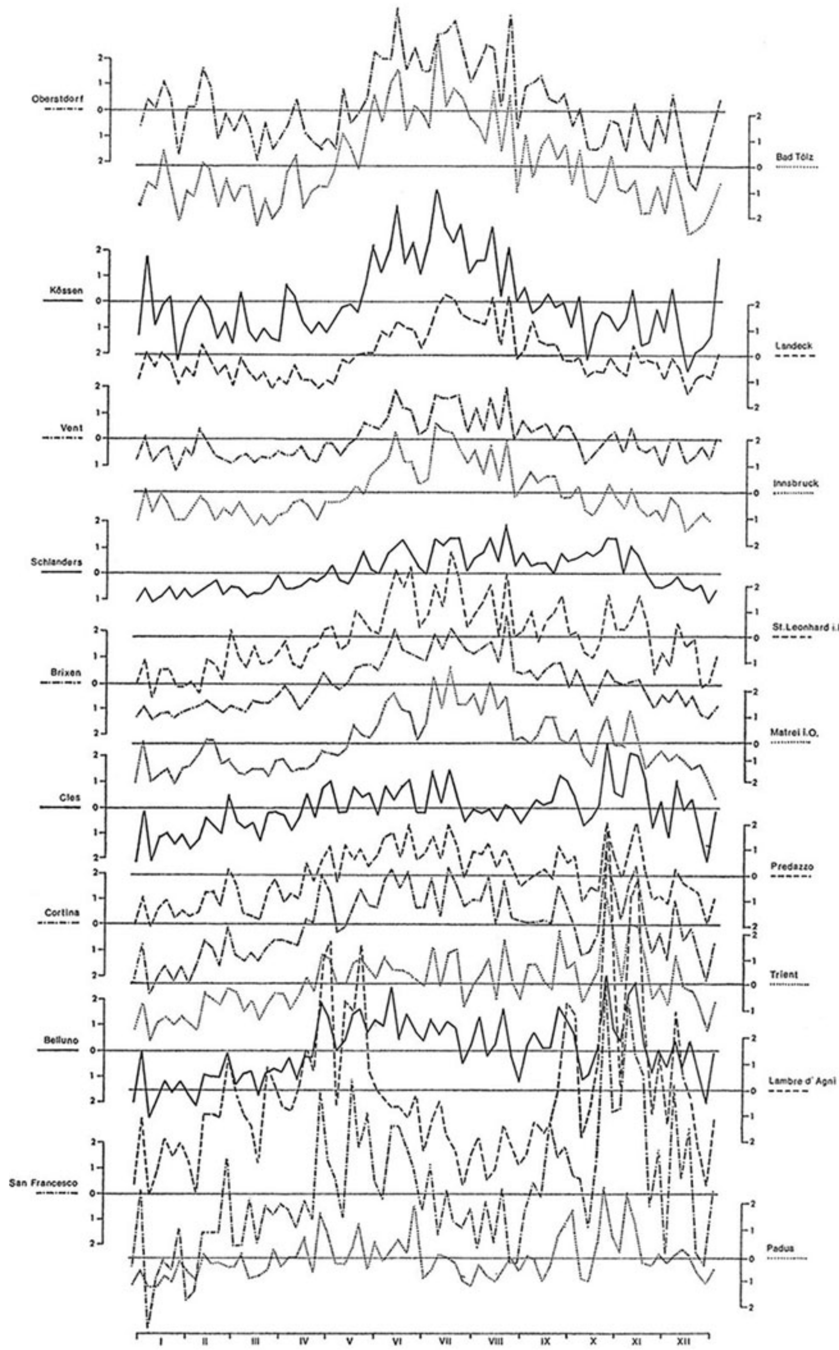

Fig. 1.4 Annual course of monthly precipitation in a cross section from N to S, approximately 10–12°E, in mm per day. Means of 1931–1960, from Fliri (1975)

1.3 Snow Cover

1.3.1 The Transition from Rain to Snow

The transition from snow to rain is expected to depend on the 0°C limit. Even at the level of formation of snow flakes it is not exactly the air temperature that determines the freezing of precipitation; it is rather the energy balance of the drop or flake, best approximated by the so called wet bulb temperature which is determined by evaporation and sublimation: at a given air temperature, drops would rather turn into snow flakes at low relative humidity. Snow forms at higher and therefore colder levels so that surface temperatures of about 1°C are an alpine-wide useful approximation for snowfall. The probability of snowfall Q may be expressed by Q = 0.6–0.1 T.

The fraction of solid vs. total precipitation depends on elevation as shown in Fig. 1.5. Considering the fairly regular dependence of temperature on elevation in this figure, it is remarkable that the fraction of solid vs. total precipitation has a much higher variance than that of temperature. The absolute values of solid precipitation F and those of total annual precipitation are dominated by their regional distribution more than by altitude.

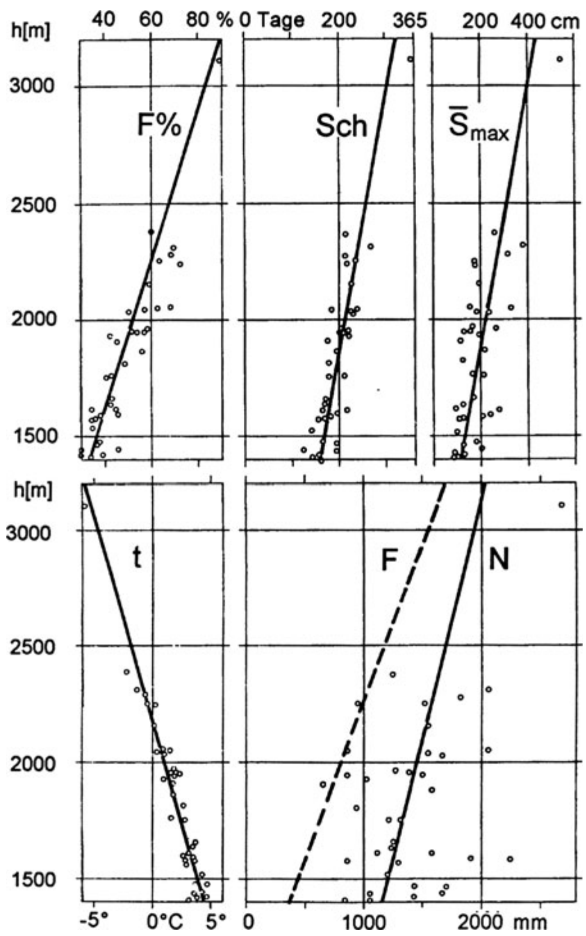

Fig. 1.5 Dependence on elevation of various quantities describing snow cover: F% fraction of solid vs. total precipitation; Sch duration of the snow cover in days; $\bar{S}_{max}$ annual means of maximum snow height; t temperature in °C; absolute values of solid F and total annual precipitation N in mm (Kuhn 1994 according to data by Lauscher)

1.3.2 Accumulation vs. Snowfall

There are several indicators of the amount of snow on the ground. Snowfall per se is best expressed as water equivalent, i.e. the height of its melt water, in mm (or kg per m^2). Fresh snow typically has a density of 100 kg m^{-3} which means that a snow cover of 10 cm has a water equivalent of about 10 mm. With a typical mean density of the winter snow cover of 300–400 kg m^{-3}, snow packs of 3 m height may be expected at elevations above the tree line; snow packs of 6 m have occasionally been observed in the Austrian Alps.

Once the snow has fallen it is generally redistributed by wind drift, also by avalanches. The amount finally lying on the ground is called accumulation and is expressed in terms of water equivalent. Wind takes snow away from ridges and crests and deposits it in concave terrain where the total accumulation may be twice as high as the original snow fall.

On a small scale, the redistribution of snow may create long lasting covers that profoundly influence vegetation. There are cornices on crests that survive into summer, and creeks that collect and preserve snow (called Schneetälchen in German literature), and there are, on the other hand, crests that are blown free of snow and may suffer much lower soil temperatures than their snow covered, insulated surroundings.

The duration of snow cover at a given elevation has been averaged for all Austrian stations and compared to the mean winter temperature at these stations. With the proper choice of scales, the two curves match very closely in Fig. 1.6.

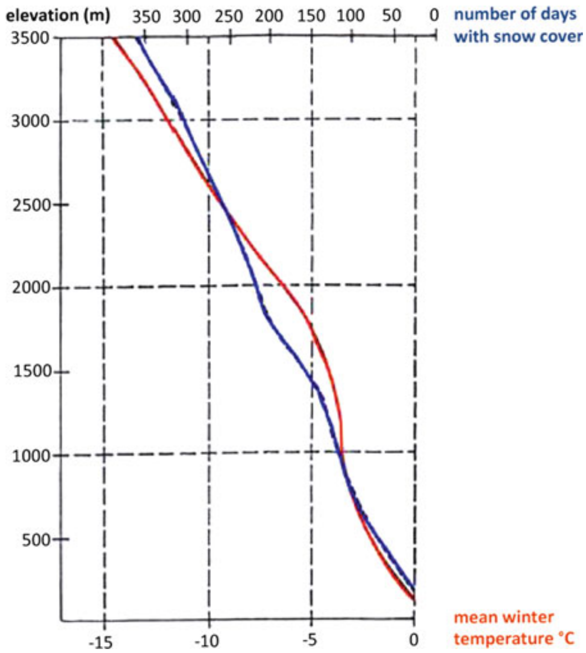

Fig. 1.6 Duration of snow cover in days and mean winter temperature vs. elevation, from Austrian stations

The seasonal development of the snow cover is determined by both accumulation and ablation. The two graphs in Fig. 1.7 show modelled snow water equivalent vs. elevation in monthly profiles from October to September for the relatively dry basin of the Rofen Valley and for the relatively humid Verwall Valley. In May at 3,050 m elevation, the snow cover in Rofen Valley is about 700 mm w.e., that in Verwall Valley is 1,400 mm w.e. From October through March both valleys have snow covers with low vertical gradients. These are determined mostly by accumulation which in turn increases above the rain/snow limit (compare the fraction of solid precipitation in Fig. 1.5), and in each month these gradients increase with elevation. In April and May, ablation starts at low elevation and diminishes the snow water equivalent, while accumulation keeps adding to the snow cover at high elevation. Thus, strong vertical gradients of snow water equivalent appear in both regions.

These two examples are mean basin values, to which local deviations to either side are caused by exposure and topography. They do, however, clearly show the natural differences in snow line elevation that exist between the dry central regions and the wet margins: Rofen Valley 3,150 m, Verwall Valley 2,550 m. In May the snowline in the early vegetation period is at 2,350 m elevation in Rofen Valley, while it is at only 1,950 m in Verwall Valley during the same period.

1.3.3 Energy and Mass Balance of the Snow Pack

The development of the snow pack is influenced by surface temperature, and hence by air temperature in two ways: temperature determines the transition from rain to snowfall; and it determines surface melting and sublimation via the energy balance of the snow (e.g. Kuhn 2008).

In the Eastern Alps, melting is the predominant form of snow ablation, it consumes an amount of 0.33 MJ kg^{-1}, subsequent evaporation requires 2.5 MJ kg^{-1}. The energy balance of a melting snow cover is

$$S\downarrow + S\uparrow + L\downarrow + L\uparrow + H + C + LS + LM = 0$$

where S is incoming and reflected solar radiation, L long wave (infrared) radiation, H turbulent sensible heat transfer, C is heat conduction in the snow, LS heat required for sublimation and LM for melting. LS includes all water that is first melted and then evaporated while LM represents the melt water that actually runs off. The fluxes S↓ and LM are not restricted to the surface but may turn over energy within the snow cover. This is due to solar radiation penetrating into the snow pack and due to melt water percolating and delivering heat to the colder interior.

Snow is a very efficient thermal insulator, which implies that strong vertical temperature gradients may exist in the snow pack. In Fig. 1.8 snow in contact with the soil remains close to 0°C all winter, while a thin top layer may cool down to about –20°C. Associated with these temperature gradients there are gradients of vapour pressure in the snow pack (saturation vapour pressure decreases by a factor of about 2 for each decrease in temperature by 10°C) which in turn lead to a transport of water vapour by diffusion in the pore space, sublimating mass from the lowest snow layers and depositing it above as so called depth hoar. This effectively changes the structure and stability of the lowest snow layers, enabling gas exchange between soil and snow.

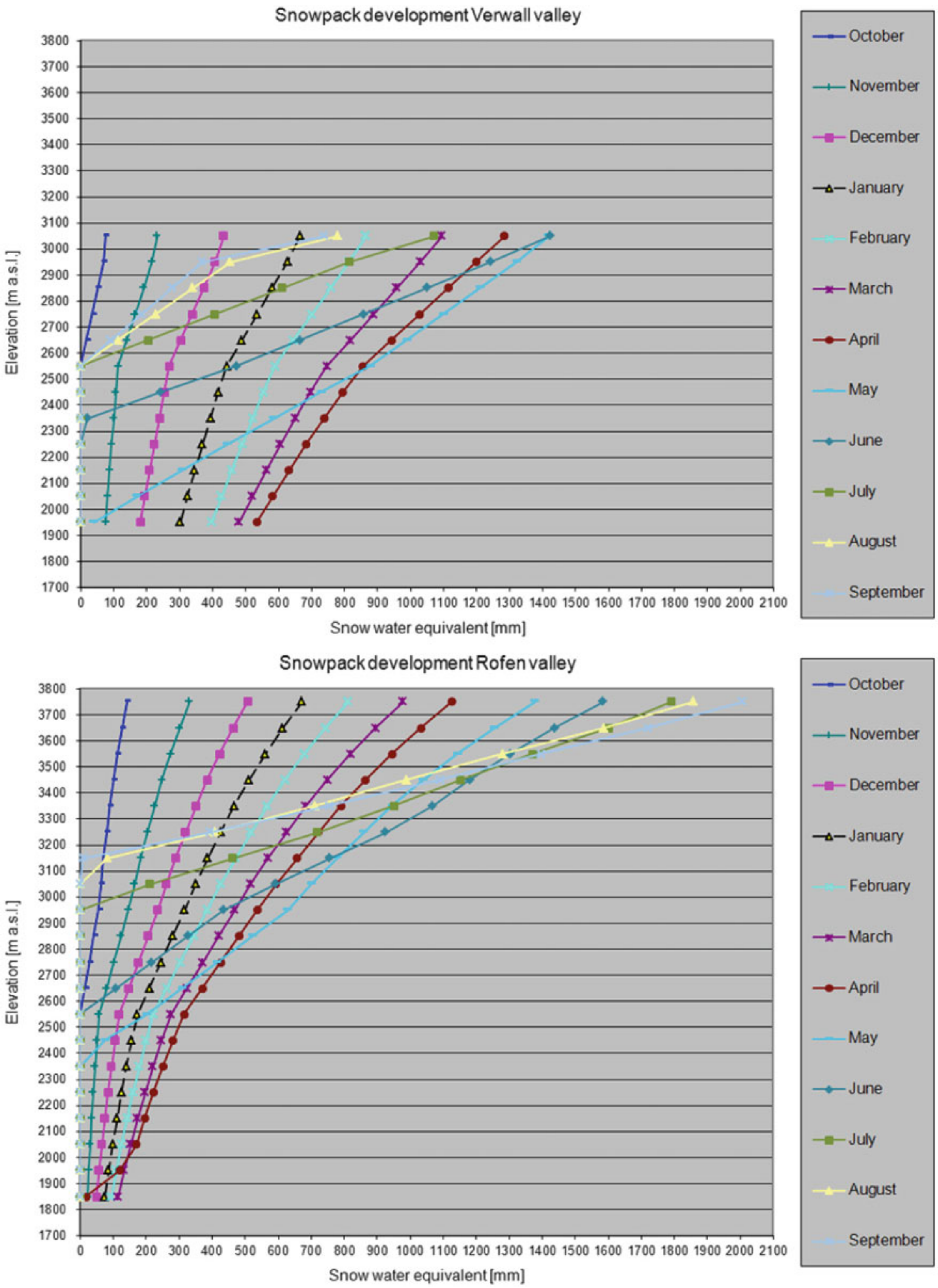

Fig. 1.7 The seasonal development of the snow cover vs. elevation modelled for two basins. Values are given in mm water equivalent. *Top*: the basin of Verwall (47.1°N, 10.2°E) with abundant precipitation, bottom: the relatively dry basin of Rofen (46.8°N, 10.8°E). Compare the peak values of snow cover at 3,000 m, and the snow line elevation in September

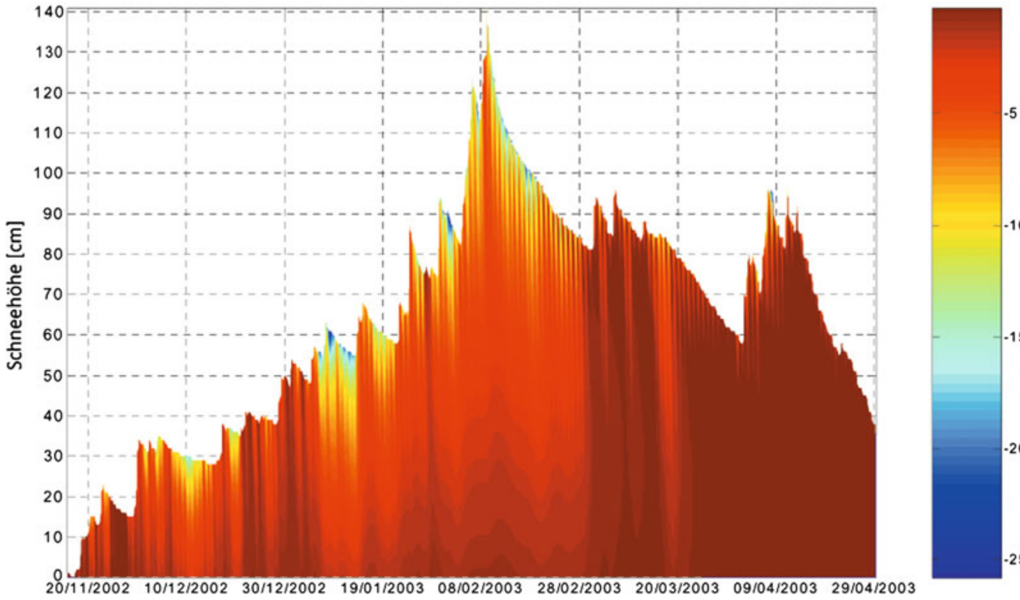

Fig. 1.8 The distribution of temperature in the snow pack at about 2,000 m elevation in the central Alps. Note the downward penetration of the daily temperature cycle and the associated phase lag. The scale on the right is in °C. From Leichtfried (2005)

1.3.4 Percolation of Rain and Melt Water Through the Snow Pack

With occasional rains in winter and with the daily melt cycle in spring, liquid water penetrates into the snow, where it soon refreezes during the accumulation period, forming ice lenses or horizontal layers that may impede vertical gas diffusion. In spring the melt water front will progressively penetrate deeper and will finally reach the soil within a few days. The percolation of melt water was modelled in Fig. 1.9 which is identical to the snow pack in Fig. 1.8.

1.3.5 Microscale Contrasts in a Broken Snow Cover

At elevations of 2,000–3,000 m global solar radiation may reach peak values of 500 W m^{-2} in early spring and 1,000 Wm^{-2} in June. This is usually more than sufficient to melt the snow which then has a surface temperature of 0°C. Dark, low albedo objects protruding from the snow, like rocks, trees or patches of bare ground, may then absorb so much solar radiation that in spite of their cold surroundings they may reach exceptionally high temperatures at a small local scale.

An example is given in Fig. 1.10 which displays the record of surface temperature of a rock of 2 m diameter extending half a meter above the snow. With a low albedo of 20% (compared to 70% of the snow), and lacking any energy loss by evaporation, it reached a surface temperature in excess of 37°C in the afternoon. Similar values have been observed on the lower parts of trees. In both examples heat is stored into the night and is transferred to the surrounding snow, soon creating bare patches around the trees or the rocks.

1.4 Acid Deposition at High Altitude

Aerosol particles and ions reach the alpine regions by dry deposition, such as dust from local sources or from long distances like the Saharan desert, or by wet deposition in rain and snow. Even if the source strength at far away places remained constant, the deposition in the Alps would always be controlled by both advection and convection, i.e. by synoptic conditions and by local atmospheric stability.

For the Alps, sources of air pollution are in the NW and in the industrial areas of northern Italy. In the typical series of synoptic events that was described in Sect. 1.2.1, it is the Eastern Alps that receive more wet deposition than the Western end of the Alps (Nickus et al. 1998).

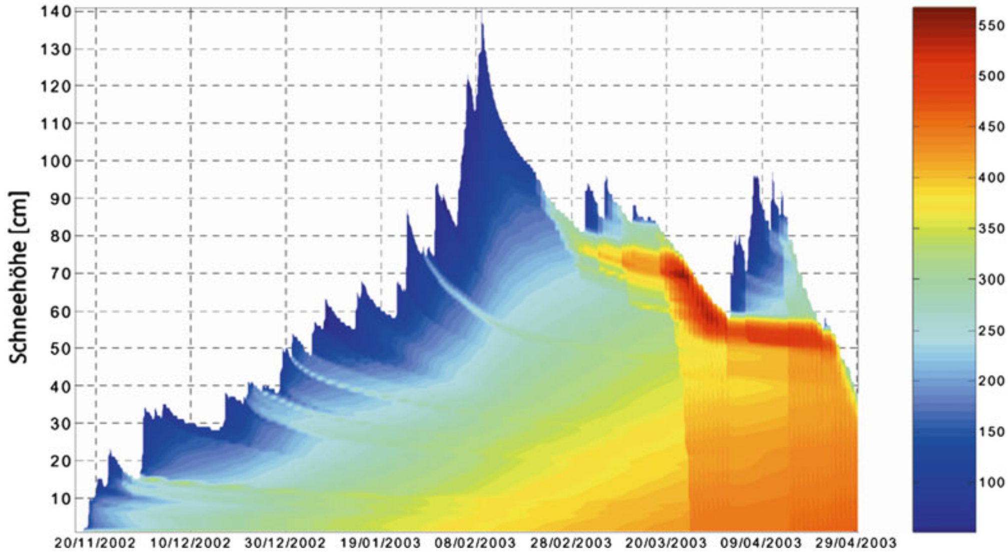

Fig. 1.9 Liquid water content of the snow pack, given as parts per thousand of the pore space on the right-hand scale. The site is identical to that in Fig. 1.8. From Leichtfried (2005)

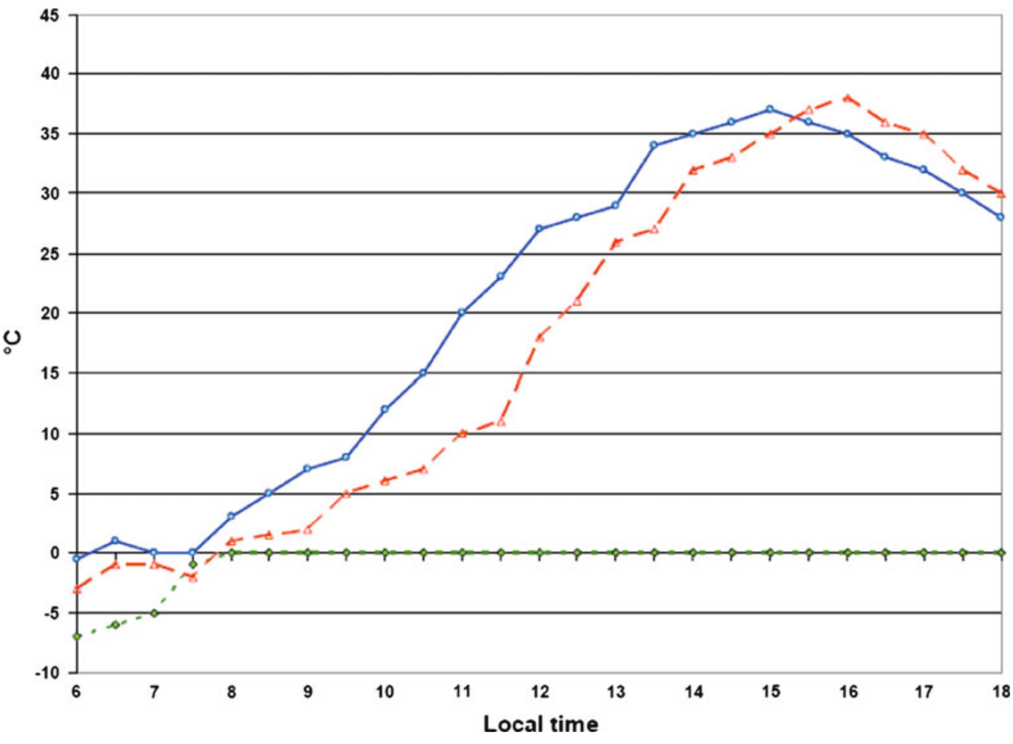

Fig. 1.10 Records of surface temperature of a rock of 2 m diameter, 0.5 m height, surrounded by snow on a clear summer day at 3,030 m. *Green curve*: snow temperature, blue south face of the rock, red west face

The seasonal development of ion concentration (μequivalents per L) and of total deposition (μequivalents per m^2) in the high alpine snow pack is generally characterized by low values in winter due to both low source strength with large areas of Europe being snow-covered, and strong atmospheric stability in the temperature inversions of alpine valleys (Kuhn et al. 1998). With the disappearance of the low land snow cover and the onset of convection in alpine regions in May and June, concentrations of sulfate, nitrate and ammonium at high altitude more than double and total loads increase by more than a factor of five considering the increase in concentration and the simultaneous increase in monthly precipitation (Fig. 1.4).

The physical and chemical processes that redistribute ions in the seasonal snow pack have been reviewed by Kuhn (2001). In the typical daily melt-freeze cycles shown in Figs. 1.8 and 1.9, ions are to a great part excluded in the process of refreezing and concentrate in the intergranular films of solution. Each day the solute becomes more concentrated and the remaining snow grains are purified. When the first melt water penetrates to the ground and runs off, it carries ions at a concentration that may be fivefold the concentration in the unmelted snow pack (Johannessen and Henriksen 1978). In the recent warming, substances of considerable age are being released into the modern alpine ecosystem, a problem that has been reported by Nickus et al. (2010).

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