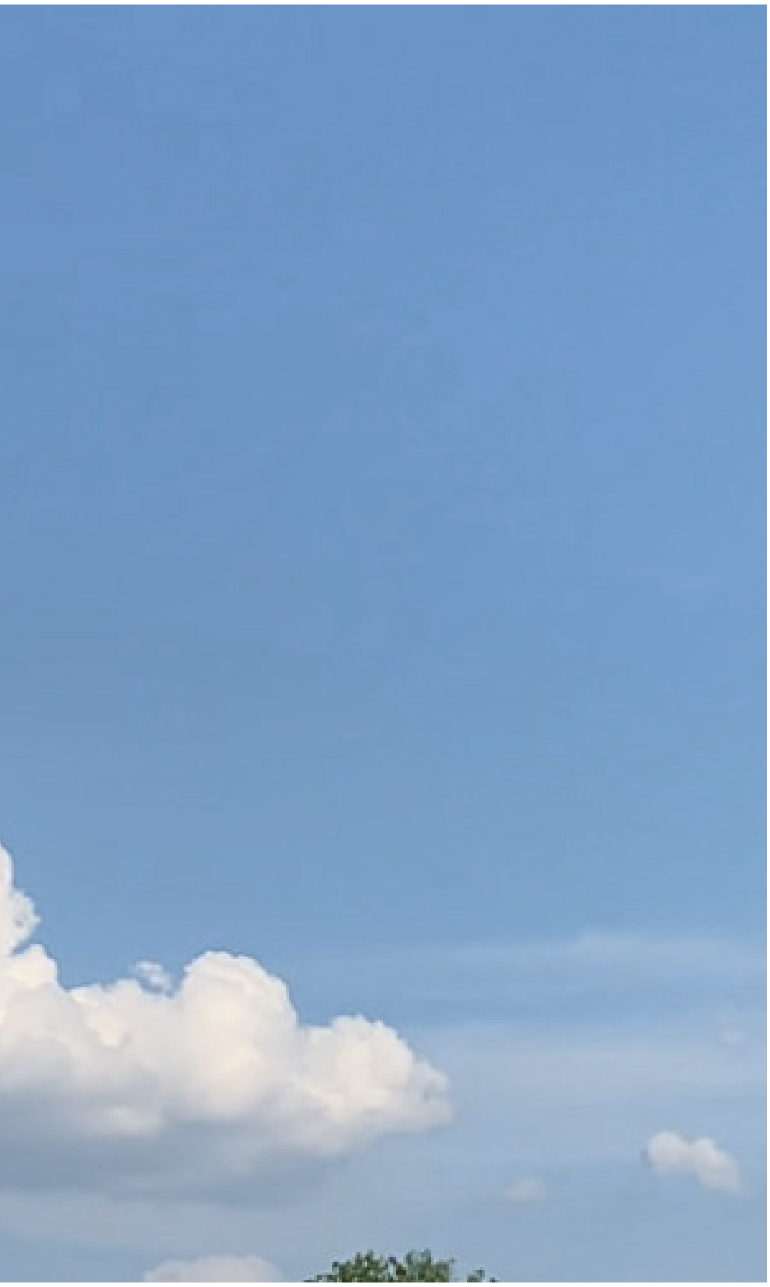
Factors influencing the

Fig 1 Cumulus congestus: durable with multiple thermal sources

thermal strength



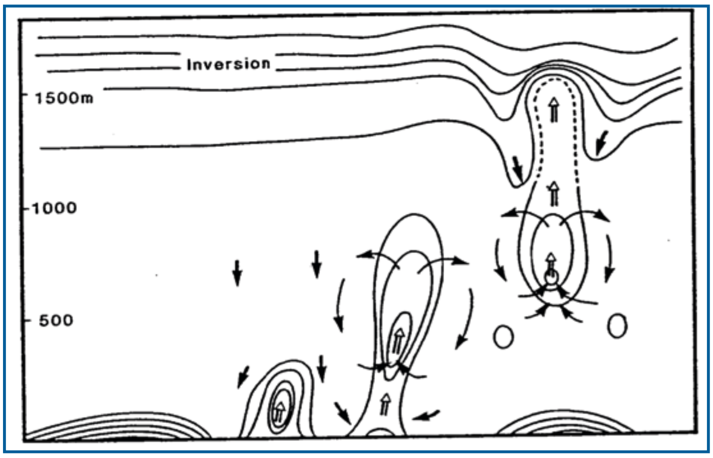
Having looked in more detail at the heating of the air near the ground and evaporation in the previous article in this series, we now want to take a closer look at the further path of heat and moisture into the atmosphere – and thus the thermal updrafts.

Authors Detlef Müller and Christoph Kottmeier

The heat (the flux of energy) first reaches the air lying directly on the ground by heat conduction of the molecular movement. This heat conduction is already replaced at low altitudes by heat transport through turbulence. Thermal turbulence, i.e. thermals with their updrafts and downdrafts, is mainly responsible for the heat transport into the higher layers. Turbulence is thus the „transporter“ of heat, but also of moisture (of water vapor!), from the air layer closest to the ground to the upper limit of the convection layer. Therefore, in meteorology these energy flows are also called „turbulent flux of sensible heat“ and the „turbulent flux of latent heat“ connected with the water vapor transport.
Key questions are how the energy flows are composed in the updrafts (distribution on heat and moisture transport), how much heat energy is supplied to the atmosphere of a region (rise of the convection layer) and how they are distributed spatially (trigger points of the updrafts).

What happens „below“ in the over-adiabatic layer?
As described in detail in the last article, the incident solar energy leads to heating and evaporation - depending on the angle of incidence, albedo (i.e. how much radiation a surface absorbs and how much it reflects), moisture content of the soil and the transpiration of the vegetation. In this process, the density of the air near the ground can decrease to the point where one reaches the threshold where lighter air is below heavier air (for a temperature-only consideration, at a temperature gradient of -3.4 °C/100 m). This is also the temperature decrease at which mirages can occur. When this threshold is reached locally, immediate vertical rearrangements (autoconvection) occur, the superadiabatic layer resting on the earth‘s surface breaks up, and an updraft develops. The air from the surrounding area, which has not yet been sufficiently warmed, flows to the breakup point and feeds the updraft – depending on the size of the warm air reservoir. Even if this critical limit is not reached, but the stratification is sufficiently unstable, thermals can be triggered. This occurs either through ever-present small eddies with dynamic lifting up to a few 10 m altitude, where the environment is cooler. Or locally, when lighter air on the ground is next to heavier air, so that it experiences lift.
How do we now arrive at the sufficient horizontal density gradient in the air near the ground? It is clear that both a higher temperature and a higher humidity can lead to a density decrease and thus to density differences. At typical summer temperatures and normal pressure, the density decrease for a relative humidity increase of about 40 % is equivalent to that for a temperature increase of 1 degree Celsius. If we had a site locally with a high evaporation rate, there would be a lack of the necessary warming to bring about a sufficient density difference in total. Thus, as a rule, the warming is the decisive factor, with evaporation – and thus moisture – making an additional contribution! The known measurements at the Leipzig TV tower show such a purely thermal breakup of the superadiabatic layer at a horizontal temperature gradient of more than 2 °C.
However, the breakup of the over-adiabatic layer can also occur „prematurely“ due to other effects, e.g. in weak wind and/or relatively homogeneous terrain due to disturbances such as a combine harvester in a field, a winch launch or even sinking cooler air. In moving air (wind), the overadiabatic layer breaks up, for example, at breakaway edges such as forests, rivers, or lakes, or at slope edges. Here, the sources may be locally fixed (trigger points like edges; more or less pulsating thermal sources), „random in nature“ (combines, sinking air), or migrating with the flux (more likely in flatter terrain). The necessary density differences for free ascent are now forced, e.g. by lifting the air parcel into higher air layers. The stronger the airflow/wind, the more likely updrafts will be triggered by disturbances in the overadiabatic layer, such as at obstacles, and the less likely air parcels will be locally „overheated“.

And how does it continue above?
Thermal packages rise – cooling dry-adiabatically in the core – up to the condensation level or a stable air layer/inversion. Compensating for the updrafts, air sinks from altitude and warms up dry-adiabatically. At the edges of the ascending and descending air „packets“, ambient air is laterally mixed in („entrainment“), causing the properties of the ascending and ambient air to converge at the edges (***Fig. 2***).
The rising air also takes along the water vapor from the layers near the ground, and due to the dry-adiabatic cooling, their relative humidity increases until the condensation level is reached. Air descending from the cumulus cloud layer or from an inversion above is usually drier. Its relative humidity decreases even further as it descends due to dry-adiabatic heating. Thus, below the cumulus condensation level, we often find much drier air in sinking areas.
As an overall result, the thermally mixed layer below the cumulus clouds and above the overadiabatic layer shows a rather uniform temperature profile with roughly one degree Celsius per 100m temperature decrease („indifferent or adiabatic stratification“). However, more detailed observations show a decrease of somewhat more in the lower third, and somewhat less than 1 °C/100 m in the upper 2/3 below the base (slightly stable stratification). Also, a weak inversion is often found at the upper boundary of the thermally dry mixed layer – or in the case of shallow cumuli in the base region. It occurs locally



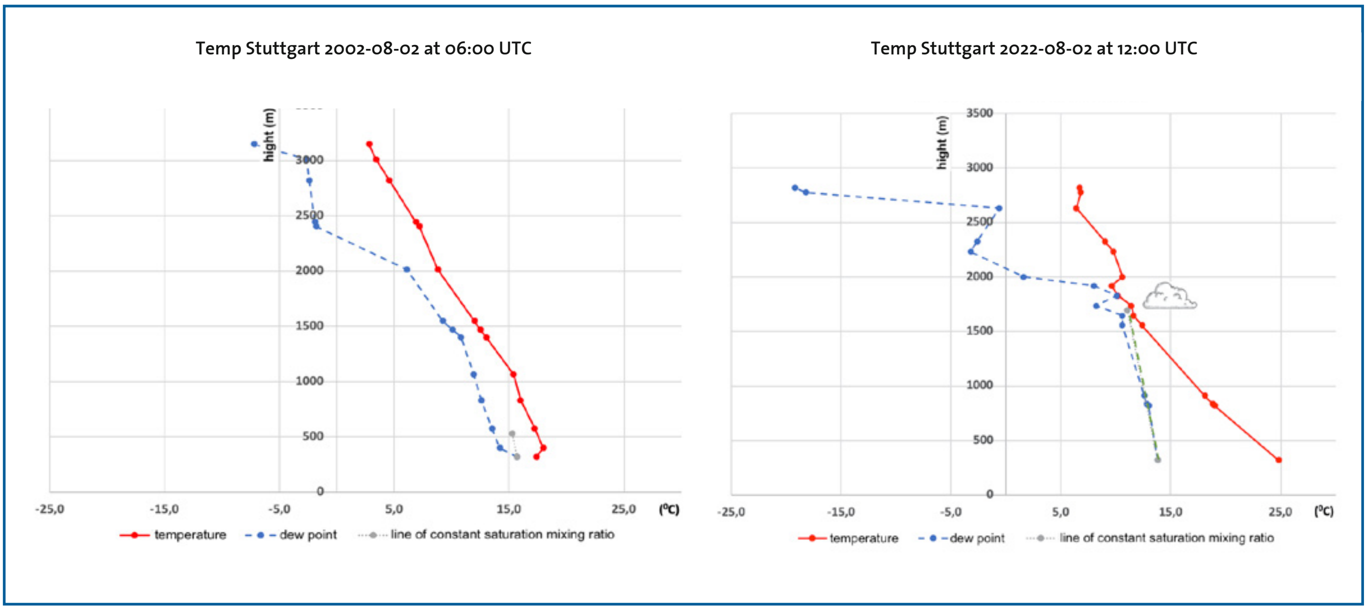


Fig 2 (above) Detachment of a thermal bubble and its penetration into a stable layer (or inversion). The solid lines are isolines of the potential temperature (OSTIV publication) **Fig 3 (bottom)** The temperature and humidity distribution in the lower 3000m altitude on 02.08.2022 over Stuttgart at 06:00 UTC and 12:00 UTC. Gray dashed is the line of equal saturation ratio starting from the dew point at the ground. In the lower third of the convection layer the air is slightly unstable (very unstable only in the lowest 50 m, overadiabatic layer), in the upper 2/3 slightly stable

due to the updrafts moving into the more stable high-altitude air.

The air humidity profile also shows a uniform course with time due to the mixing processes: Below the cumulus condensation level, the dew point curve roughly follows the mean saturation mixing ratio of the thermally mixed layer. In the case of greater moisture supply at the ground/greater evaporation, the moisture towards the ground is thereby slightly increased, and especially in the case of greater dryness above the thermal ceiling, the dew point curve is slightly inclined „upwards to the left“ (with altitude towards lower dew points). ***Fig. 3*** shows an example of the change of the vertical temperature and humidity profile over Stuttgart on 02.08.2022 in the course of the day.

Interesting insights into what happens in the thermally mixed boundary layer are provided by the glider measurements of Ronald Niederhagen. Ronald performed a series of temperature and humidity surveys with his Ventus in 2018 and 2019 and published results together with Oliver Predelli (OSTIV Congress 2019 „Correct interpretation of glider in-flight environmental sensing in thermal updrafts“). Ronald gave me the opportunity to view and further process his data.

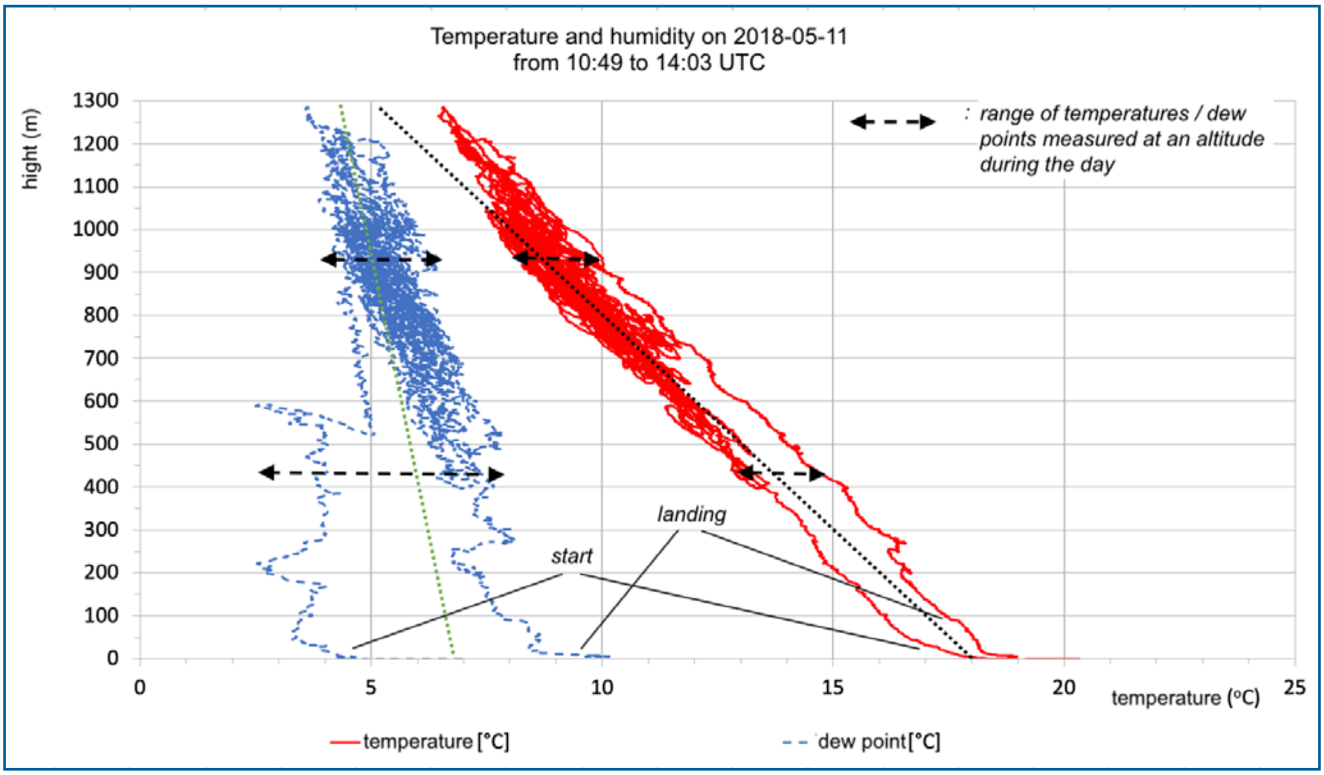


Fig 4 Flight measurement of temperature and humidity on 11.05.2018 by Ronald Niederhagen.
4 a (top) All temperature and dew point measurement data. Marked is the variance of temperature and dew point over the course of the day, caused by the diurnal warming of the air as well as variations of measured values both in the used/flown through updrafts and in the surrounding air partly associated with sinking
4 b (bottom) Measurement data of the takeoff and landing. The temperature profile is nearly dry-adiabatic. The pronounced lability measured near the ground during takeoff is partly due to the previous heating of the temperature sensor on the ground, so it should not be overestimated. In the upper part of the landing approach, the stratification is slightly dry-stable. While the dew point curves during the F-tow as well as during the landing approach show clearly fluctuating values (turbulent mixing of the thermal boundary layer!), the curve in the first upwind at higher dew point runs parallel to the line of equal saturation mixing ratio. The dew point on the ground was 7 °C, which fits well with the intersection of the line with the airfield level. During the landing approach, the dew point curve averaged shows a course with almost constant saturation mixture ratio

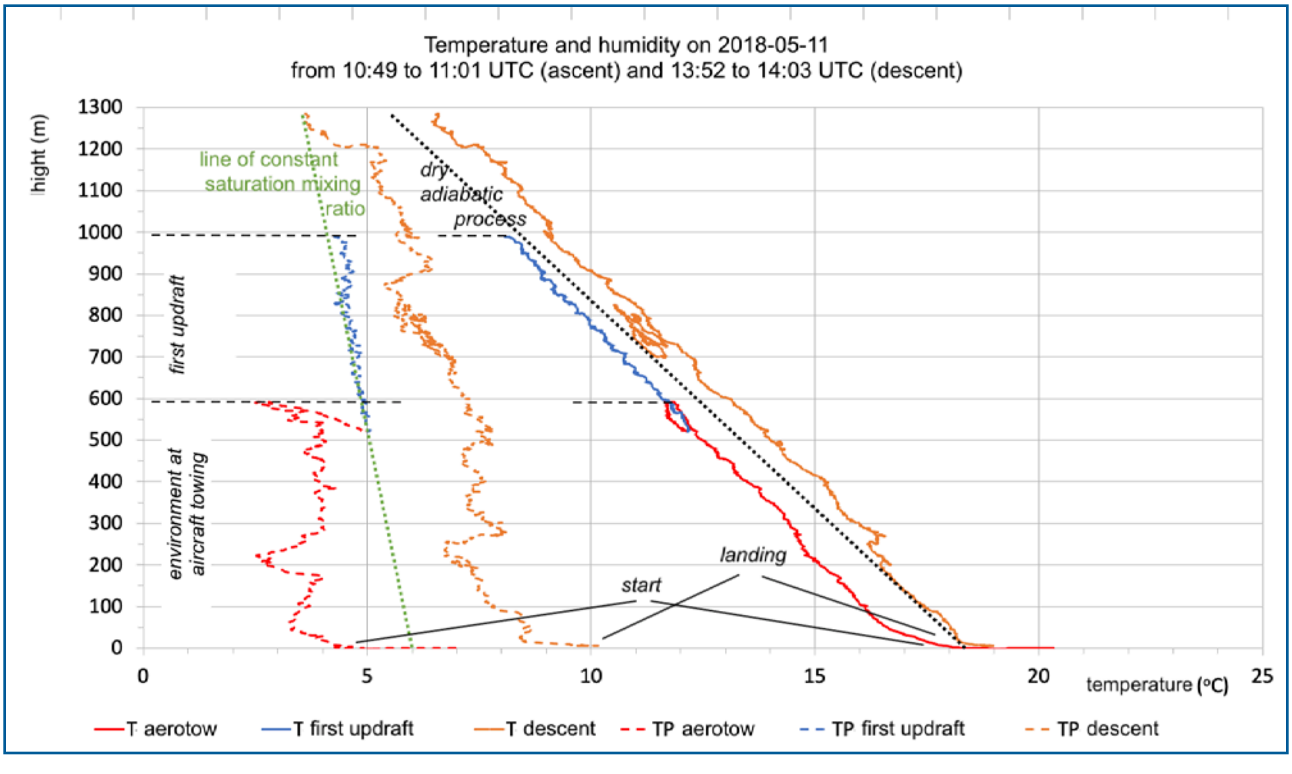


Fig. 4a and b show the graphical processing of the temperature and humidity measurements on 11.05.2018 and confirm quite well the statements about the temperature and humidity distribution in the convection space below the base during the course of the day: the growth of the nearly dry-adiabatic mixed layer below the cumulus base, the over-adiabatic layer and slightly stable stratification at altitude, and with respect to of the humidity the nearly constant saturation mixing ratio – however with sometimes more and sometimes less pronounced fluctuations: especially in very dry high-altitude air, regular „downbursts“ with spatially limited extension and low humidity were flown through in the vicinity of updrafts.
The often observed rapid decrease of the excess temperature in updrafts compared to their surroundings can best be explained by the merging of smaller updrafts or the collection of smaller updrafts by single larger ones. This results in a mixing of updrafts with lower and stronger excess temperatures, and to a large extent in a complete reduction of the excess temperature at half the boundary layer height. The updrafts have then reached their typical climb values. Further acceleration can then only occur through the buoyancy of the water vapor component of the air and through active clouds (see below) above the updrafts.

What happens during cumulus cloud formation

Cumulus clouds form when the core of the updraft has reached the cumulus condensation level (CCN). Roland Stull (1993) refers to „forced clouds“ that initially form. Active updrafts have caused air to rise high enough so that the CCN has just been reached. If the surrounding air of the cumulus cloud is stratified with moisture stability, the cloud air can continue to rise at an accelerated rate. This is called an „active cloud“. If the supply of thermal air, and thus the further supply of moisture, is interrupted, the dissipating cloud is called a „passive cloud“.
There are days when only forced clouds exist, each as long as it can maintain itself in a moist adiabatic stable environment as a result of further moisture supply from an active updraft. The cumuli that form often have a small vertical and horizontal extent, frayed appearance, no sharp base, and are short-lived (keywords: „drawn blue thermal“). The stronger the rise and the more prolonged the supply of warm, humid thermal air to a cumulus cloud, the more moisture can condense and the sharper the cloud base is formed.
Active cumulus clouds are more long-lived. The additional acceleration of the rising air within the cloud causes a „suction effect“ below the cumulus, which increases in its effect with cloud thickness. Here we then also observe the thermal sources moving along with the flux. This can be observed extremely with local thunderstorms, which „follow“ the warm air roughly in the direction of the flux, but are also deflected, e .g., at rivers (***Fig. 1 on the first double page***). Or in ordered thermals: the cumuli of the cloud streets are regularly distributed and migrate with the flux.

How the Earth‘s Surface Affects Cumulus Clouds

Regions with similar characteristics in terms of land surface and soil type typically show differences in land use, reflectivity, and moisture. Accordingly, different land use areas with different soil moisture adopt different temperatures and humidity resulting in different CCNs. This can be the case on different days, but also simultaneously at different locations in a region. Our computational approach differs from that used for temp analysis, which is usually performed for only one characteristic temperature and dew point temperature.
As a result, there is an altitude range of possible CCNs depending on the temperature and dew point of the air near the ground that forms a cumulus cloud as an updraft. For example, over moist surfaces, more of the solar energy is used for

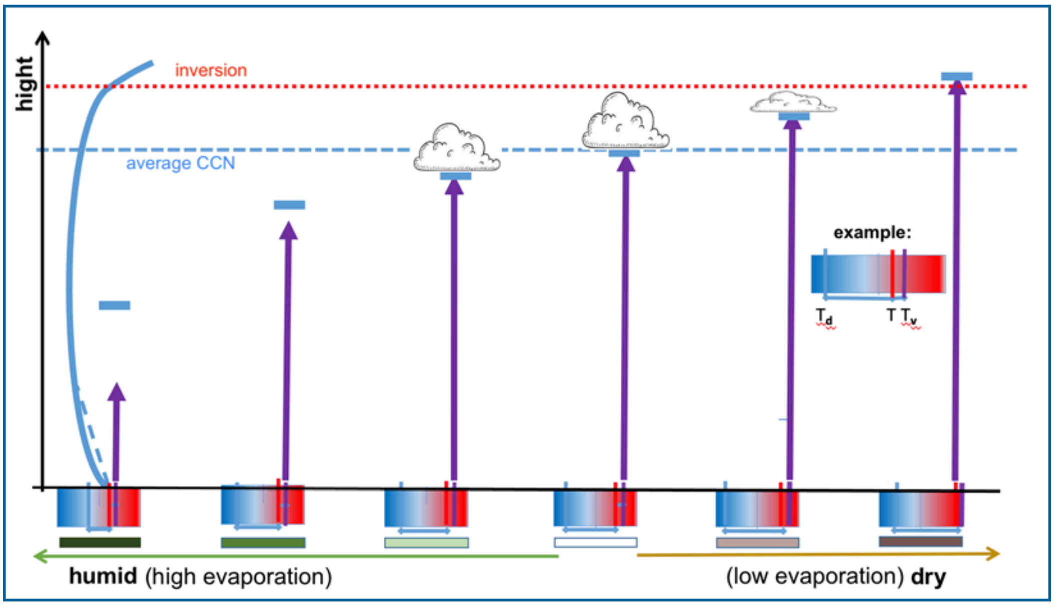


Fig. 5 Behavior of thermal packages lifting off from near the ground with increasing temperature and decreasing moisture on the ground. In the case of heterogeneous earth surfaces, simultaneous updrafts sometimes reach lower or higher heights (own calculations, sketch after Stull, 1993)

evaporation, leaving less for heating. As a result, the initial lift is relatively low and, because of the higher humidity, the CCN is also lower. ***Fig 5*** and the accompanying ***Table 1*** illustrate this.

Table 1 shows six updrafts that can all occur simultaneously somewhere in an area with a heterogeneous surface. Due to their different temperature and humidity near the ground as a result of drier or wetter surfaces, they have different convection ceilings and condensation levels. Of course, in cooler and more humid regions, updrafts can also occur: the radiative energy supplied has to go somewhere! However, the lift caused by temperature and humidity differences may be too low to reach the CCN: we then experience weak climbing in cloud-free space during the preflight. If, on the other hand, neighboring rising air reaches 100 % relative humidity, a Cu forms. The higher the temperature at constant dew point and thus the dew point difference, the higher the base will be. If the rising air is very dry near the ground in other updrafts (right profile), this can lead to blue thermals with an updraft top above the CCN in the vicinity during the course of the day.

The effects of a strongly fluctuating moisture supply on the ground – and thus a strongly fluctuating ground-level humidity – can be clearly observed on the day after local showers: here, even in regions with relatively homogeneous ground conditions, differences in elevation in the CCN of more than 200m occur. But also notoriously thermally weak points in the landscape and the hot spots with frequent triggering can be explained in this way. The stronger the airflow / wind, the more likely updrafts are triggered by disturbances in the superadiabatic layer, the less air parcels are locally „overheated“ as well as moisture differences are spatially balanced and the more uniform the CCN is.

In order to take into account the effect of humidity on air density for buoyancy considerations, the „virtual temperature“ was determined for the pairs of values: it is the temperature that dry air would have to have in order to have the same density as moist air at a lower real temperature and the same pressure. It becomes very clear how relatively small the so-called „virtual surcharge“ due to a higher air humidity is compared to an increase in temperature. Near the ground, the factor determining the buoyancy is the temperature, but at altitude it is usually the humidity! Temperature AND humidity at ground level determine the CCN.

Let us summarize

The superadiabatic layer allows air to rise in the form of updrafts. However, it also acts as a filter that only allows air parcels to ascend when there is sufficient density difference or uplift as a result of disturbances. Density differences are temperature and humidity dependent: in thermal initiation temperature differences are decisive, in higher altitude humidity differences. Stronger energy input at ground level – e.g., due to lower albedo – causes increased heating, which locally increases the absorption capacity for water vapor and thus evaporation. Ground-level temperature differences are then also associated with moisture differences.

The stronger the wind, the less the possibility of locally larger

Table 1 Pairs of values for the cases of different thermal updrafts shown in Fig. 5 with increasing temperature and decreasing humidity of the thermal packages lifting off from near the ground

Cases	T_s (°C)	T_d (°C)	Spread (°C)	Rel. humidity	CCN (m)	e (hPa)	q (kg/kg)	Virtual surcharge (K)
1	**22**	**17,2**	4,8	74,2 %	600	19,67	0.0123	2,21
2	**23**	**16,6**	6,4	67,2 %	800	18,94	0,0119	2,14
3	**24**	**16,0**	8,0	60,9 %	1000	18,23	0,0114	2,06
4	**25**	**15,4**	9,6	55,2 %	1200	17,54	0,0110	1,99
5	**26**	**14,8**	11,2	50,0 %	1400	16,88	0,0106	1,93
6	**27**	**14,2**	12,8	45,4 %	1600	16,23	0,0102	1,86

heating differences building up, and the more uniform the thermal conditions. Near the ground and at larger obstacles, however, the thermal is then more disrupted. On thermal days, updrafts and downdrafts cause turbulent upward transport of sensible and latent heat by releasing/absorbing heat and water vapor with the environment via a lateral exchange. Initially higher ground-level temperature differences between surrounding air and rising air parcel are quickly reduced by the coalescence of smaller updrafts or collection of smaller updrafts by single larger ones.

The turbulent processes provide a typical vertical temperature and humidity profile below the cumulus condensation level: the downdrafts are also dryadiabatic to a first approximation. Due to the constant mixing between updrafts and downdrafts, the ambient air is almost dry-adiabatically stratified. The moisture distribution approaches the constant mean saturation mixing ratio of the air in the thermally mixed layer below the cumulus condensation level.

The vertical velocity of an updraft

„Archimedian lift" is a common model of thought, according to which motion in convection is explained by forces. The buoyant force is considered separately from the much larger pressure gradient force, which is formed by the mass buildup of the air at rest and opposes gravity. The buoyancy acceleration $a = g * \Delta r/r$ (with g gravity acceleration, r density and Δr density difference), is determined by the density difference that arises in the atmosphere due to temperature and humidity differences. Accordingly, the equation is often summarized in the form $a = g * \Delta T_V/T_V$, where T_V is the virtual temperature. If the air was enabled to rise near the ground by buoyancy, it will continue to rise without further acceleration, so all that is needed is an initial impulse. Even when entering a slightly stable stratification of the atmosphere, buoyancy remains effective until the parcel assumes the density of its surroundings. Only downward acceleration due to greater density than the environment will stop the updraft.

Observations show typical ΔT_V of up to 3 degrees near the ground. This corresponds to a boyant force of about 0.1 m/s^2, 1/100 of the acceleration due to gravity. If the ascent begins from rest, the rate of ascent increases to 1 m/s within 10 sec and then continues beyond that. The distance traveled upward is calculated as $h = ½ a t^2$. After 10 sec, the air has risen 5 m in height, and after 30 sec it has risen to a height of 45 m with a vertical velocity of 3 m/s (theoretically, without considering flux resistances). Since this means that the superheated layer near the ground has been left, the acceleration can only continue at a reduced rate. Even a ΔT_V of 0.3 or 0.5 degrees is quite sufficient to explain the observed vertical speeds of up to 5 m/s.

Various aircraft measurements show that the temperature differences between rising and surrounding air decrease with altitude. Measurements by Carsten Lindemann, for example, showed an average of only 0.3K at an altitude of 200m! From about 1/3 of the thermally mixed layer the updrafts reach an environment with slightly stable stratification, with which then finally the temperature differences disappear or even become negative. Then the growing humidity difference to

Fig. 6 **Active cumulus cloud with migrating thermal source (1 to 4) and triggering of a new updraft on its windward side. T he mentioned phenomena around cumulus cloud formation can be found under the QR code on the right in various YouTube videos**

the environment as well as the existing kinetic energy of the updraft provide for the further climb.

The cumulus cloud - an essential factor

In our article „Cumulus clouds – flywheels and guideposts, sometimes also spoilers of thermals“ (issue 01-2023, QR code end of article) we presented, among other things, measurement results on the distribution of the climb. Thus, in blue thermals a maximum of the vertical climb is usually located about in the middle of the convection area, in cloud thermals at half base height with a second maximum in pronounced cumulus clouds. The vertical velocity maximum at half base height suggests that the virtual temperature between rising and surrounding air is equalized by this height and its influence on the density difference is then negative. The rising air then initially continues to move despite negative acceleration solely due to its inertia. It is easy to estimate that in a thermal column of e.g. 500 m height and a mean diameter of 400 m an air mass of about 60000 tons(!) is moved, comparable to the kinetic energy of an ICE train at 100 km/h. This movement must first be slowed down!

In active cumulus clouds – i .e. cumuli that form in a moisture-stable environment – the additional acceleration of the rising air in the cumulus cloud (moist adiabatic cooling in a moist adiabatic stable environment) also causes a „suction effect" below the cloud. In sum with the increasing humidity difference to the environment, this explains the usually increasing climb below the base of well-developed cumulus clouds, but also the longer lifetime and the better mean climb. In addition, at somewhat higher wind speeds, one observes that new updrafts and clouds are triggered on the windward side of clouds by the associated updrafts (***Fig. 6***).

Thus, a number of factors are decisive for the vertical speed of updrafts and the distribution of climb with altitude:

- the initial density difference near the ground (temperature difference! Initial acceleration. Pronounced with incoming cold air!)
- the size of the heat reservoir at ground level (how much warmer air can flow in)
- the moving air volume (height AND diameter)
- the humidity distribution in the ambient air
- the cloud base
- the temperature and humidity distribution in the cloud level (cloud thickness; moist-stable/instable, inversions)/ the temperature and humidity distribution of the stable air layer in which the climb comes to a halt at the end.

When we speak of a „average climb rate" in gliding, we mean the average value of the glider's encountered flyable climb! As shown above – and as experience teaches – there is always a significantly weaker climb in the atmosphere, whereby the associated updrafts usually do not reach the condensation level. The range of climb in updrafts in the atmosphere can be quite considerable!

The current models for gliding weather forecasts use models for the vertical distribution of the temperature difference between rising and surrounding air and from this a corresponding distribution of the climb over the convection space to calculate an average climb value. These then also take into account the penetration of the updrafts into more stable air layers at altitude. A simple table (***below***) for estimating the mean (flyable) climb rate on a thermal day (see „Handbook of meteorological forecasting for soaring flight" of WMO/ OSTIV) still provides sufficiently good values for the mean flyable climb rate on a thermal day.

In arid climates (Australia, South Africa) the numerical values have to be multiplied by a factor of 1.5 to 2. There, the convection area reaches high altitudes and the amplification effect of small Cu-clouds is no longer pronounced. ♦

	max. height of the dry adiabatic layer	Flyable average climb rate
Blue thermals	1 km	1,0 m/s
	2 km	2,0 m/s
	3 km	3,0 m/s
Thermal with small Cu	1 km	1,2 m/s
	2 km	2,4 m/s
	3 km	3,6 m/s
Thermals with heavier Cu clouds and persistent cold-air advection	1 km	1,5 m/s
	2 km	3,0 m/s
	3 km	4,5 m/s

References:

WMO/OSTIV: „Handbook of meteorological forecasting for soaring flight", OSTIV-Projekt 2000
Xiao-Ming Hu And Ming Xue: „The Use of High-Resolution Sounding Data to Evaluate and Optimize Nonlocal PBL Schemes for Simulating the Slightly Stable Upper Convective Boundary Layer", Monthly Weather Review, 2019
Stull, Robert B.: „Boundary-Layer Cumulus Over Land: Some Observations and Conceptual Models" ECMWF Workshop on Boundary-Layer Clouds. Reading, 1993.

Additional info:
„Cumulus clouds – flywheels and guideposts, sometimes spoilers of thermals".