

Chapter 2

Wind Regimes

The principle origin of the winds in the Earth's atmosphere and the potentially available power from these winds have been qualitatively described in [Sect. 1.4](#). This general description of the driving forces for the wind has to be brought into a mathematical formulation for precise turbine load and energy yield calculations and predictions. Therefore, this chapter will present the basic wind laws in the free atmosphere. Vertical wind profiles in atmospheric boundary layers over different surface types will be presented in the subsequent [Chaps. 3–5](#).

2.1 Global Circulation

Flow patterns and winds emerge from horizontal surface and atmospheric temperature contrasts on all spatial scales from global to local size. Globally, the tropical belt and the lower latitudes of the Earth are the main input region for solar energy, while the higher latitudes and the poles are the regions with a negative energy balance, i.e. the Earth here loses energy through thermal radiation. Ocean currents and atmospheric heat conduction are not sufficient to compensate for this differential heating of the globe. The global atmospheric circulation has to take over as well. Main features of this global atmospheric circulation are the Hadley cell, the Ferrel cell and the polar cell which become visible from a latitude-height plot showing an average over all longitudes of the winds in the troposphere and stratosphere. The Hadley cell exhibits a direct thermal circulation. Warm air rises near the equator, moves towards the poles aloft and descends in the subtropics. The region of sinking motion is characterised by large anticyclones in the surface pressure field and deserts. Likewise, the polar cell exhibits a direct thermal circulation as well. Here, cold air sinks over the poles and rises at higher latitudes. This is the reason for generally high pressure over the poles. In between the Hadley cell and the polar cell lies the thermally indirect Ferrel cell. This cell is characterised by rising colder air at higher latitudes and sinking warmer air in the

subtropics. This circulation is indirect and it is the result of the integral effect over all the moving cyclones in this belt of temperate latitudes. Effectively, the Ferrel cells transport warmer air towards the poles near the ground and colder air towards the tropics aloft. This indirect circulation is maintained by energy conversions from potential energy into kinetic energy in the moving cyclones of the temperate latitudes.

The just described system of cells would only produce meridional winds, i.e. winds from South to North or vice versa. The Earth's rotation is modifying this meridional circulation system by the Coriolis force. Winds towards the poles get a westerly component, winds towards the equator an easterly component. Therefore, we mainly observe westerly winds at the ground in the Ferrel cell while we observe easterly winds at the ground in the Hadley cell and the polar cell. The north easterly winds near the ground of the Hadley cell are also known as the trade winds. These global wind cells have a spatial scale of roughly 10,000 km. The global wind system is modified by the temperature contrasts between the continents and the surrounding oceans and by large north-south orientated mountain ranges, in particular those at the west coasts of the Americas. These modifications have a spatial scale of some 1,000 km. Even smaller land-sea wind systems in coastal areas may have an order of 100 km; mountain and valley wind systems can be even smaller in the order of several tens of kilometres. All these wind systems may be suitable for wind power generation.

While the trade winds and the winds in the polar cell exhibit quite some regularity and mainly have seasonal variations, the winds in the Ferrel cell are much more variable in space and time. Near-surface wind speeds in normal cyclones can vary between calms and about 25 m/s within a few hours. Wind speeds in strong hibernal storms of the temperate latitudes can reach about 35–40 m/s while wind speeds in subtropical hurricanes easily reach more than 50 m/s. Cut-off wind speeds of modern wind energy turbines are between 25 and 30 m/s. Thus strong storms in temperate latitudes may lead to phases where the wind potential can no longer be used. These hibernal storms are most likely in Northwestern Europe, Northeastern Canada, the Pacific coasts of Canada and Alaska as well as the southern tips of South America, Africa and Australia.

Hurricanes are called typhoons in Southeast Asia and cyclones in India. The occurrence of hurricanes can even threaten the stability of the construction of the turbines, because they can come with wind speeds above those listed in the IEC design standards. The hurricane risks have been investigated by Rose et al. (2012). In particular, the planning of offshore wind parks in hurricane-threatened areas needs special attention. According to the map of natural hazards published by the reinsurance company Munich Re, hurricane-prone areas are the southern parts of the Pacific coasts and the Atlantic coasts of the United States and Central America, Eastern India and Southeast Asia, Madagascar and the northern half of Australia.

There are very strong winds on even smaller scales such as thunderstorm downbursts, whirlwinds and tornados, but their variability and destructive force is

not suited for windpower generation. Rather turbines have to be constructed in a way that they can stand these destructive forces while being shut off. See also Sects. 2.6 and 6.5 for wind hazards.

2.2 Driving Forces

The equations in the following Subchapters describe the origin and the magnitude of horizontal winds in the atmosphere. We will start with the full set of basic equations in Sects. 2.2.1 and 2.2.2 and will then introduce the usual simplifications which lead to the description of geostrophic and gradient winds in Sect. 2.3. Geostrophic and gradient winds, which blow in the free atmosphere above the atmospheric boundary layer, have to be considered as the relevant external driving force in any wind potential assessment and any load assessment. Vertical variations in the geostrophic and gradient winds are described by the thermal winds introduced in Sect. 2.4.

2.2.1 Hydrostatic Equation

The most basic explanation of the wind involves horizontal heat gradients. The sun heats the Earth's surface differently according to latitude, season and surface properties. This heat is transported upward from the surface into the atmosphere mainly by turbulent sensible and latent heat fluxes. This leads to horizontal temperature gradients in the atmosphere. The density of air, and with this density the vertical distance between two given levels of constant pressure, depends on air temperature. A warmer air mass is less dense and has a larger vertical distance between two given pressure surfaces than a colder air mass. Air pressure is closely related to air density. Air pressure is a measure for the air mass above a given location. Air pressure decreases with height. In the absence of strong vertical accelerations, the following hydrostatic equation describes this decrease:

$$\frac{\partial p}{\partial z} = -g\rho = -\frac{gp}{RT} \quad (2.1)$$

where p is air pressure, z is the vertical coordinate, g is the Earth's gravity, ρ is air density, R is the specific gas constant of air, and T is absolute air temperature. With typical near-surface conditions ($T = 293$ K, $R = 287$ J kg⁻¹ K⁻¹, $p = 1,000$ hPa and $g = 9.81$ ms⁻²) air pressure decreases vertically by 1 hPa each 8.6 m. In wintry conditions, when $T = 263$ K, pressure decreases 1 hPa each 7.7 m near the surface. At greater heights, this decrease is smaller because air density is

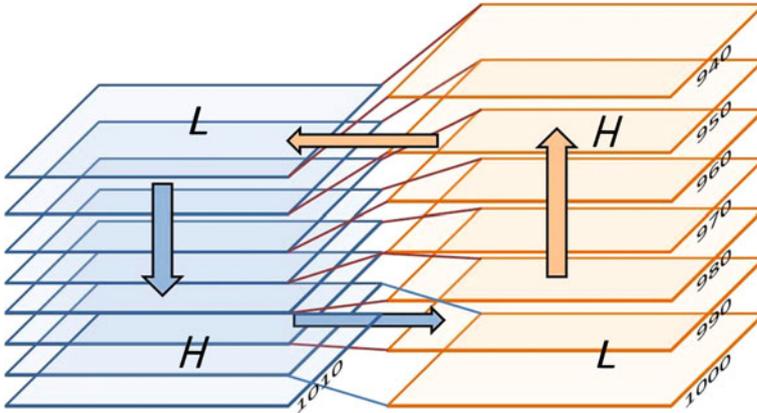


Fig. 2.1 Vertical pressure gradients in warmer (*right*) and colder (*left*) air. Planes symbolize constant pressure levels. Numbers give air pressure in hPa. Capital letters indicate high (*H*) and low (*L*) pressure at the surface (*lower letters*) and on constant height surfaces aloft (*upper letters*). *Arrows* indicate a thermally direct circulation

decreasing with height as well. At a height of 5.5 km the air pressure is at about half of the surface value, and thus, the pressure only decreases by 1 hPa every 15 m. An (unrealistic) atmosphere at constant near-surface density would only be 8 km high!

The consequence of (2.1) is that the pressure in warm air masses decreases more slowly with height than in cold air masses. Assuming a constant surface pressure, this would result in horizontal pressure gradients aloft. A difference in 30° in air mass temperature will cause a 1.36 hPa pressure gradient between the warm and the cold air mass 100 m above ground. This pressure gradient produces compensating winds which tend to remove these gradients. In reality, surface pressure sinks in the warmer region (“heat low”). This situation is depicted in Fig. 2.1. In a situation with no other acting forces (especially no Coriolis forces due to the rotating Earth) this leads to winds blowing from higher towards lower pressure. Such purely pressure-driven winds are found in land-sea and mountain-valley wind systems. This basic effect is depicted in term III in the momentum budget equations that will be introduced in the following section.

2.2.2 Momentum Budget Equations for the Wind

A mathematical description of the winds is most easily done by considering the momentum balance of the atmosphere. Momentum is mass times velocity. The momentum budget equations are a set of differential equations describing the

Table 2.1 Latitude-dependent Coriolis parameter f in s^{-1} for the northern hemisphere. The values in both columns are negative for the southern hemisphere

Latitude (in degrees)	Coriolis parameter in s^{-1}
30	0.727×10^{-4}
40	0.935×10^{-4}
50	1.114×10^{-4}
60	1.260×10^{-4}

acceleration of the three wind components. In complete mass-specific form, they read (mass-specific means that these equations are formulated per unit mass, the mass-specific momentum has the physical dimension of a velocity. Therefore, we say wind instead of momentum in the following):

$$\frac{\partial u}{\partial t} + \vec{v} \cdot \nabla u + \frac{1}{\rho} \frac{\partial p}{\partial x} - fv + f^*w \mp v \frac{|\vec{v}|}{r} + F_x = 0 \quad (2.2)$$

$$\frac{\partial v}{\partial t} + \vec{v} \cdot \nabla v + \frac{1}{\rho} \frac{\partial p}{\partial y} + fu \pm u \frac{|\vec{v}|}{r} + F_y = 0 \quad (2.3)$$

$$\frac{\partial w}{\partial t} + \vec{v} \cdot \nabla w + \frac{1}{\rho} \frac{\partial p}{\partial z} - g - f^*u + F_z = 0 \quad (2.4)$$

I II III IV V VI VII

where u is the wind component blowing into positive x direction (positive in eastward direction), v is the component into y direction (positive in northward direction) and w is the vertical wind (positive upward). The wind vector is $\vec{v} = (u, v, w)$, the horizontal Coriolis parameter is $f = 2\Omega \sin\varphi$ where Ω is the rotational speed of the Earth and φ is the latitude (see Table 2.1), the vertical Coriolis parameter is $f^* = 2\Omega \cos\varphi$, r is the radius of curvature, and F_x, F_y , and F_z are the three components of the frictional forces, which will be specified later. The Eqs. (2.2)–(2.4), which are called Eulerian equations of motion in meteorology, are a special form of the Navier-Stokes equations in hydrodynamics.

Term I in Eqs. (2.2)–(2.4) is called inertial or storage term, it describes the temporal variation of the wind components. The non-linear term II expresses the interaction between the three wind components. Term III specifies the above-mentioned pressure force. Term IV, which is present in (2.4) only, gives the influence of the Earth’s gravitation. Term V denotes the Coriolis force due to the rotating Earth. Term VI describes the centrifugal force in non-straight movements around pressure maxima and minima (the upper sign is valid for flows around lows, the lower sign for flows around high pressure systems). The last term VII symbolizes the frictional forces due to the turbulent viscosity of air and surface friction.

The terms in (2.2)–(2.4) may have different magnitudes in different weather situations and a scale analysis for a given type of motion may lead to discarding

some of them. Nearly always, the terms containing f^* are discarded because they are very small compared to all other terms in the same equation. In larger-scale motions term VI is always neglected as well. Term VI is only important in whirl winds and close to the centre of high and low pressure systems. Looking at the vertical acceleration only (Eq. (2.4)), terms III and IV are dominating. Equating these two terms in (2.4) leads to the hydrostatic equation (2.1) above.

There is only one driving force in Eqs. (2.2–2.4): the abovementioned pressure force which is expressed by term III. The constant outer force due to the gravity of the Earth (term IV) prevents the atmosphere from escaping into space. The only braking force is the frictional force in term VII. The other terms (II, V, and VI) just redistribute the momentum between the three different wind components. Thus, sometimes terms V and VI are named “apparent forces”. In the special case when all terms II to VII would disappear simultaneously or would cancel each other perfectly, the air would move inertially at constant speed. This is the reason why term I is often called inertial term.

2.3 Geostrophic Winds and Gradient Winds

The easiest and most fundamental balance of forces is found in the free troposphere above the atmospheric boundary layer, because frictional forces are negligible there. Therefore, our analysis is started here for large-scale winds in the free troposphere. The frictional forces in term VII in Eqs. (2.2–2.4) can be neglected above the atmospheric boundary layer. Term VI is also very small and negligible away from pressure maxima and minima. The same applies to term II for large-scale motions with small horizontal gradients in the wind field. A scale analysis shows that the equilibrium of pressure and Coriolis force is the dominating feature and the inertial term I can be neglected as well. This leads to the following two equations:

$$-\rho f u_g = \frac{\partial p}{\partial y} \quad (2.5)$$

$$\rho f v_g = \frac{\partial p}{\partial x} \quad (2.6)$$

with u_g and v_g being the components of this equilibrium wind, which is usually called geostrophic wind in meteorology. The geostrophic wind is solely determined by the large-scale horizontal pressure gradient and the latitude-dependent Coriolis parameter, the latter being in the order of 0.0001 s^{-1} (see Table 2.1 for some sample values). Because term VII had been neglected in the definition of the geostrophic wind, surface friction and the atmospheric stability of the atmospheric boundary layer has no influence on the magnitude and direction of the geostrophic wind. The modulus of the geostrophic wind reads:

$$|v_g| = \sqrt{u_g^2 + v_g^2} \quad (2.7)$$

The geostrophic wind blows parallel to the isobars of the pressure field on constant height surfaces. Following Eqs. (2.5) and (2.6), a horizontal pressure gradient of about 1 hPa per 1,000 km leads to a geostrophic wind speed of about 1 m/s. In the northern hemisphere, the geostrophic wind blows counter-clockwise around low pressure systems and clockwise around high pressure systems. In the southern hemisphere the sense of rotation is opposite.

Term VI in Eqs. (2.2–2.4) is not negligible in case of considerably curved isobars. The equilibrium wind is the so-called gradient wind in this case:

$$-\rho f u = \frac{\partial p}{\partial y} \pm \frac{\rho u |\vec{v}|}{r} \quad (2.8)$$

$$\rho f v = \frac{\partial p}{\partial x} \mp \frac{\rho v |\vec{v}|}{r} \quad (2.9)$$

Once again, the upper sign is valid for flows around lows, the lower sign for flows around high pressure systems. The gradient wind around low pressure systems is a bit lower than the geostrophic wind (because centrifugal force and pressure gradient force are opposite to each other), while the gradient wind around high pressure systems is a bit higher than the geostrophic wind (here centrifugal force and pressure gradient force are unidirectional).

Sometimes, in rare occasions, the curvature of the isobars can be so strong that the centrifugal force in term VI is much larger than the Coriolis force in term V so that an equilibrium wind forms which is governed by pressure forces and centrifugal forces only. This wind, called cyclostrophic wind by meteorologists, is found in whirl winds and tornados.

The geostrophic wind and the gradient wind are not height-independent in reality. Horizontal temperature gradients on levels of constant pressure lead to vertical gradients in these winds. The wind difference between the geostrophic winds or gradient winds at two different heights is called the thermal wind.

2.4 Thermal Winds

We introduced in Sect. 2.3 the geostrophic wind as the simplest choice for the governing large-scale forcing of the near-surface wind field. The geostrophic wind is an idealized wind which originates from the equilibrium between pressure gradient force and Coriolis force. Until now we have always anticipated a barotropic atmosphere within which the geostrophic wind is independent of height, because we assumed that the horizontal pressure gradients in term III of (2.2) and (2.3) are independent of height. This is not necessarily true in reality and the

deviation from a height-independent geostrophic wind can give an additional contribution to the vertical wind profile as well. The horizontal pressure gradient becomes height-dependent in an atmosphere with a large-scale horizontal temperature gradient. Such an atmosphere is called baroclinic and the difference in the wind vector between geostrophic winds at two heights is called thermal wind. The real atmosphere is nearly always at least slightly baroclinic, thus the thermal wind is a general phenomenon.

Thermal winds do not depend on surface properties. So they can appear over all surface types addressed in [Chaps. 3–5](#).

Differentiation of the hydrostatic equation (2.1) with respect to y and differentiation of the definition equation for the u -component of the geostrophic wind (2.5) with respect to z leads after the introduction of a vertically averaged temperature T_M to the following relation for the height change of the west–east wind component u :

$$\frac{\partial u}{\partial z} = -\frac{g}{fT_M} \frac{\partial T_M}{\partial y} \quad (2.10)$$

Subsequent integration over the vertical coordinate from the roughness length z_0 to a height z gives finally for the west–east wind component at the height z :

$$u(z) = u(z_0) - \frac{g(z - z_0)}{fT_M} \frac{\partial T_M}{\partial y} \quad (2.11)$$

The difference between $u(z)$ and $u(z_0)$ is the u -component of the thermal wind. A similar equation can be derived for the south–north wind component v from Eqs. (2.1) and (2.6):

$$v(z) = v(z_0) + \frac{g(z - z_0)}{fT_M} \frac{\partial T_M}{\partial x} \quad (2.12)$$

Following (2.10) and (2.11), the increase of the west–east wind component with height is proportional to the south–north decrease of the vertically averaged temperature in the layer between z_0 and z . Likewise, (2.12) tells us that the south–north wind component increases with height under the influence of a west–east temperature increase. Usually, we have falling temperatures when travelling north in the west wind belt of the temperate latitudes on the northern hemisphere, so we usually have a vertically increasing west wind on the northern hemisphere.

Equations (2.11) and (2.12) allow for an estimation of the magnitude of the vertical shear of the geostrophic wind, i.e. the thermal wind from the large-scale horizontal temperature gradient. The constant factor $g/(fT_M)$ is about 350 m/(s K). Therefore, a quite realistic south–north temperature gradient of 10^{-5} K/m (i.e., 10 K per 1,000 km) leads to a non-negligible vertical increase of the west–east wind component of 0.35 m/s per 100 m height difference.

The thermal wind also gives the explanation for the vertically turning winds during episodes of cold air or warm air advection. Imagine a west wind blowing from a colder to a warmer region. Equation (2.12) then gives an increase in the

south–north wind component with height in this situation. This leads to a backing of the wind with height. In the opposite case of warm air advection the wind veers with height.

2.5 Boundary Layer Winds

The wind speed in the atmospheric boundary layer must decrease to zero towards the surface due to the surface friction (no-slip condition). The atmospheric boundary layer can principally be divided into three layers in the vertical. The lowest layer which is only a few millimetres deep is laminar and of no relevance for wind energy applications. Then follows the surface layer (also called constant-flux layer or Prandtl layer), which may be up to about 100 m deep, where the forces due to the turbulent viscosity of the air dominate, and within which the wind speed increases strongly with height. The third and upper layer, which usually covers 90 % of the boundary layer, is the Ekman layer. Here, the rotational Coriolis force is important and causes a turning of the wind direction with height. The depth of the boundary layer usually varies between about 100 m at night with low winds and about 2–3 km at daytime with strong solar irradiance.

Scale analysis of the momentum Eqs. (2.2–2.4) for the boundary layer show the dominance of terms III, V, and VII. Sometimes, for low winds in small-scale motions and near the equator, the pressure force (term III) is the only force and a so-called Euler wind develops, which blows from higher pressure towards lower pressure. Such nearly frictionless flows rarely appear in reality. Usually an equilibrium between the pressure force and the frictional forces (terms III and VII) is observed in the Prandtl layer, and an equilibrium between the pressure force, the Coriolis force and the frictional forces (terms III, V, and VII) is observed in the Ekman layer. The Prandtl layer wind is sometimes called antitriptic wind. No equation for the antitriptic winds analog to (2.5, 2.6) or (2.8, 2.9) is available, since neither term III nor term VII contains explicitly the wind speed.

The Prandtl layer is characterised by vertical wind gradients. The discussion of Prandtl layer wind laws which describe these vertical wind speed gradients is postponed to Chap. 3. The vertical gradients are much smaller in the Ekman layer, so that it is meaningful to look at two special cases of (2.2) and (2.3) in the following section.

In a stationary Ekman layer the terms III, V, and VII balance each other, because term I vanishes. This layer is named from the Swedish physicist and oceanographer W. Ekman (1874–1954), who for the first time derived mathematically the influence of the Earth's rotation on marine and atmospheric flows. A prominent wind feature in the Ekman layer is the turning of wind direction with height.

The vertical profiles of these boundary layer winds over different surface types will be analysed in more detail in the upcoming [Chaps. 3–5](#).

2.6 Thunderstorm Gusts and Tornados

There are strong winds which cannot be used for wind energy generation, because they are short-lived and rare in time and place, such that their occurrence is nearly unpredictable. Most prominent among these phenomena are thunderstorm gusts and tornados. Offshore tornados are called waterspouts. They can be so violent that they can damage wind turbines. Therefore, the probability of their occurrence and their possible strength should be nevertheless investigated during the procedure of wind turbine siting.

While onshore tornados mostly form in the afternoon and the early evening at cold fronts or with large thunderstorms when surface heating is at a maximum, offshore waterspouts are more frequent in the morning and around noon when the instability of the marine boundary layer is strongest due to nearly constant sea surface temperatures (SST) and cooling of the air aloft overnight ([Dotzek et al. 2010](#)). However, the seasonal cycle is different. Onshore tornados most frequently occur in late spring and summer. Offshore waterspouts peak in late summer and early autumn. In this season, the sea surface temperature of shallow coastal waters is still high, while the first autumnal rushes of cold air from the polar regions can lead to an unstable marine boundary layer favourable for waterspout formation ([Dotzek et al. 2010](#)).

Although the characteristics of tornado formation are understood in principle today, the prediction of their actual occurrence remains difficult because a variety of different favourable conditions have to be met simultaneously. In general, following [Houze \(1993\)](#) and [Doswell \(2001\)](#), tornado formation depends largely on the following conditions:

- (potential) instability with dry and cold air masses above a boundary layer capped by a stable layer preventing premature release of the instability;
- a high level of moisture in the boundary layer leading to low cloud bases;
- strong vertical wind shear (in particular for mesocyclonic thunderstorms);
- pre-existing boundary layer vertical vorticity (in particular for non-mesocyclonic convection).

A rough estimation how often a tornado could hit a large wind park is given in [Sect. 6.5](#).

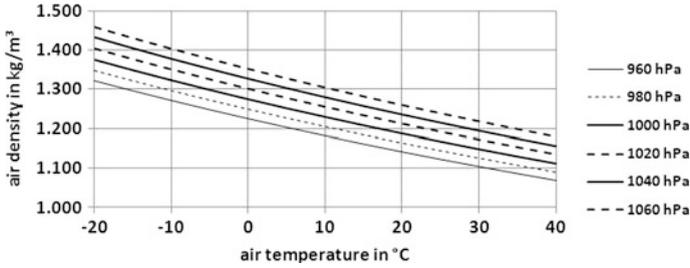


Fig. 2.2 Near-surface air density as function of air temperature and surface pressure

2.7 Air Density

Apart from wind speed, the kinetic energy content of the atmosphere also depends linearly on air density (see Eq. (1.1)). Near-surface air density, ρ is a direct function of atmospheric surface pressure, p and an inverse function of air temperature, T . We have from the state equation for ideal gases:

$$\rho = \frac{p}{RT} \quad (2.13)$$

where $R = 287 \text{ J kg}^{-1} \text{ K}^{-1}$ is the universal gas constant. Equation (2.13) is equivalent to the hydrostatic equation (2.1) above. Figure 2.2 shows air density for commonly occurring values of surface temperature and surface pressure. The Figure illustrates that air density can be quite variable. A cold wintertime high pressure situation could easily come with a density around 1.4 kg/m^3 , while a warm low pressure situation exhibits an air density of about 1.15 kg/m^3 . This is a difference in the order of 20 %.

Figure 2.2 is valid for a dry atmosphere. Usually the atmosphere is not completely dry and the modifying effect of atmospheric humidity has to be considered. Humid air is less dense than completely dry air. Meteorologists have invented the definition of an artificial temperature which is called virtual temperature. The virtual temperature, T_v is the temperature which a completely dry air mass must have in order to have the same density as the humid air at the actual temperature, T . The virtual temperature is defined as:

$$T_v = T(1 + 0.609q) \quad (2.14)$$

where q is the specific humidity of the air mass given in kg of water vapour per kg of moist air. The temperatures in Eq. (2.14) must be given in K. The difference between the actual and the virtual temperature is small for cold air masses and low specific humidity, but can be several degrees for warm and very humid air masses. Figure 2.2 can be used to estimate air density of humid air masses, if the

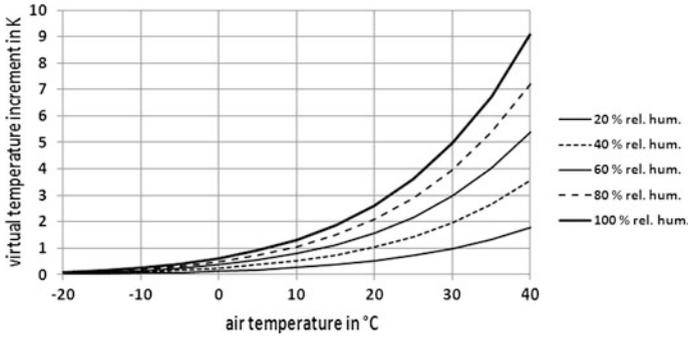


Fig. 2.3 Virtual temperature increment $T_v - T$ in K as function of air temperature and relative humidity for an air pressure of 1,013.25 hPa

temperature in Fig. 2.2 is replaced by the virtual temperature. Figure 2.3 gives the increment $T_v - T$ by which the virtual temperature is higher than the actual air temperature as function of temperature and relative humidity of the air for an air pressure of 1,013.25 hPa.

Figure 2.3 shows that the virtual temperature increment is always less than 1 K for temperatures below the freezing point, but reaches, e.g. 5 K for saturated humid air at 30 °C. The virtual temperature increment slightly decreases with increasing air pressure. A 1 % increase in air pressure (10 hPa) leads to a 1 % decrease in the virtual temperature increment. Thus, the determination of the exact density of an air mass requires the measurement of air pressure, air temperature and humidity.

Air density decreases with height, because air pressure decreases with height as given in (2.1). We get from (2.1) to (2.13) (Ackermann and Söder 2000):

$$\rho(z) = \frac{p_r}{R\bar{T}} \exp\left(\frac{-g(z - z_r)}{R\bar{T}}\right) \quad (2.15)$$

p_r is the air pressure at a reference level z_r , and \bar{T} is the vertical mean temperature of the layer over which the density decrease is computed. Temperature is decreasing with height as well; therefore equation (2.15) should only be used for small vertical intervals.

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