Chapter 6

Atmosphere

6.1 Overview

The atmosphere is a body of air that envelops the Earth. It extends up to extremely high altitudes, into what is normally thought of as space, although the vast bulk of the air mass is below 10 km. Thus, in terms of aviation (and, indeed, the entire biosphere), it is a relatively thin layer of air (when compared with the mean radius of the Earth, which is approximately 6370 km). Its overall structure is summarised in Figure 6.1. This emphasises three different ways of separating the atmosphere into layers, based on composition, temperature and radiation.

Below about 80 km, the composition of the atmosphere is an almost uniform mixture of nitrogen (~78 %), oxygen (~21 %) and argon (~1 %), with traces of other gases and variable amounts of water vapour and particulate matter. Of the other gases, carbon dioxide, ozone and methane are of particular significance in the debate on climate change. This is the so-called *homosphere*. Above about 80 km, the composition changes with altitude as the residual air tends to stratify according to the molecular mass of the component gases. Nitrogen is virtually absent above 200 km, above which the dominant components tend to be atomic oxygen (up to 1000 km), helium (up to 2000 km) and hydrogen (above 2000 km). This is the *heterosphere*.

The atmospheric layers that are defined by temperature are the *troposphere*, the *stratosphere*, the *mesosphere* and the *thermosphere*. The upper boundaries are delineated by the tropopause, stratopause, mesopause and thermopause, respectively. Note that the mesopause marks the top of the homosphere.

The troposphere contains about 90 % of the air mass and is characterised by a linear reduction in temperature with altitude, down to about −50 °C at the tropopause (at mid-latitudes). Given this very low temperature, almost no atmospheric water can exist at higher altitude. The thickness of the troposphere varies continually and seasonally. Typically, it is more than 20 km at the equator and less than 10 km at the poles. Note that most cloud formation occurs below 8 km.

The stratosphere is strongly influenced by ozone, which is created by a reaction between atomic and molecular oxygen. Atmospheric flow patterns cause a high concentration of ozone at around 25 km, which is known as the *ozone layer*. Because ozone absorbs ultraviolet radiation from the Sun, the temperature in the *upper* stratosphere (above the ozone layer) increases with altitude. In the lower stratosphere, temperature is virtually constant and circulation is limited, making this layer very stable. Particulate matter (e.g. carbon particles derived from the combustion

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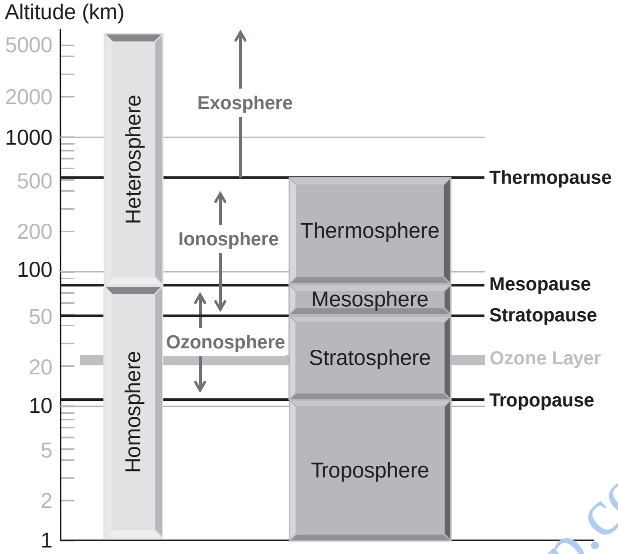

Figure 6.1 Overall structure of the atmosphere.

of petroleum products) that enters this layer is likely to remain there for prolonged periods (which is thought to be the key factor in global dimming).

The mesosphere is characterised by rapid temperature decrease with altitude, which is largely due to the role of carbon dioxide absorbing infrared radiation from the surface of the Earth (which is thought to be a key factor in global warming). It is also part of the process of ozone creation that operates in the stratosphere.

The thermosphere is a region of rapid temperature increase with altitude, leading to kinetic temperatures[1] of about 1500 K at 300 km altitude. This is the result of absorbing ultraviolet radiation from the Sun and so this varies on a daily (or *diurnal*) cycle as well as on cycles of sunspot activity. Energy cannot be reradiated at this level and so must follow a slow conduction path to lower altitudes (below 100 km altitude) from where it can be radiated at night. Note that the atmosphere at these extreme altitudes is so rarified that, while the kinetic temperature is high, the specific heat capacity is practically zero. In other words, the external environment is cold!

Considering radiation, the designated atmospheric layers are the *ionosphere*, the *ozonospere* and the *exosphere*. These are of peripheral interest to this book and so will not be described in detail. In the ionosphere, cosmic and solar radiation ionises gas molecules and also causes the dissociation of molecular oxygen into atomic oxygen. This layer has high electrical conductivity (which affects radio communication) and can be energised by the solar wind (which creates aurorae in the skies around polar latitudes). The ozonosphere is a layer that corresponds with the stratosphere and the mesosphere, in which ultraviolet radiation dissociates any stray water vapour from lower layers into hydrogen and oxygen atoms. The process of ozone creation then takes place, as discussed already. The exosphere is essentially everything outside and it extends indefinitely. This contains the magnetosphere, which marks the extent of the geomagnetic field (as discussed at length in Chapter 4).

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[1] Kinetic temperature is a measure of the energy state of individual atoms and molecules.

6.2 Standard Atmosphere Models

The development of standard models of the atmosphere has been underway since the 1920s, motivated by the need to provide a common basis for calibrating aircraft instruments and for analysing aircraft performance. In essence a standard atmosphere is a steady-state model that is averaged over a full year. It considers the air mass as a single entity that rotates with the Earth and comprises a homogeneous mixture of constituent gases (neglecting the effects of water vapour, particulate matter and solar activity). In this way, regional, diurnal and seasonal fluctuations are removed and so too is any dependency on latitude.

ESDU[2] Data Item 7702 1b offers a useful summary of developments, leading up to the two principal derivations that apply to aviation, namely the ISO[3] Standard Atmosphere[4] (ISO 2533:1975) and the US Standard Atmosphere (COESA, 1976). Early American and European efforts were harmonised in 1952 when the International Civil Aviation Organization (ICAO) adopted a standard atmosphere for altitudes up to 20 km. Various extensions followed as a result of experimental data gathered from high-altitude aircraft and, especially, rockets and satellites. Models are readily available for atmospheric properties up to 1000 km but, clearly, aviation rarely exceeds 20 km. Over this range all derivations of standard atmosphere are practically identical.[5] This includes the two mentioned already together with others, such as those adopted by ICAO (1993) and the World Meteorological Organization.

ISO 2533:1975 covers altitudes up to 80 km. The *International Standard Atmosphere* (ISA) refers strictly to the portion of this model up to 50 km, with the reminder being designated as an Interim Standard Atmosphere. In this context, the term ‘interim’ is not really appropriate as no serious change is envisaged in the foreseeable future. In any event, as already suggested, this is of no interest to aviation.

A useful extension to the ISA is ISO 5878:1982, which defines a set of reference models for aerospace use. This presents information on the temporal and spatial variations in atmospheric properties at levels between the surface and 80 km. This is achieved via reference atmospheres for tropical, subtropical, mid-latitude, subarctic and arctic regions. These are *nonstandard* or *off-standard* atmospheres that are used for more specific design investigations. This concept will be developed later in the chapter, with particular reference to ESDU Data Items 78008c and 78012b.

6.3 ISA Constants and Relationships

The International Standard Atmosphere is underpinned by the following constants:

$$\text{Mean Earth Radius} \qquad r_0 = 6356766\,\text{m} \tag{6.1}$$

$$\text{Gravitational Acceleration} \qquad g_0 = 9.80665\,\text{m.s}^{-2} \tag{6.2}$$

$$\text{Speed of Sound} \qquad a_0 = 340.294\,\text{m.s}^{-1} \tag{6.3}$$

$$\text{Pressure} \qquad p_0 = 1.01325 \times 10^{-5}\,\text{Pa} \tag{6.4}$$

$$\text{Temperature} \qquad T_0 = 288.15\,\text{K} \tag{6.5}$$

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[2] Engineering Sciences Data Unit (ESDU).

[3] International Standardisation Organisation (ISO).

[4] The coordination body within ISO is TC6 (*Technical Committee, Aircraft and Space Vehicles*) SC6 (*Sub-Committee, Standard Atmosphere*).

[5] A common definition exists for the altitude range from –2 km up to 32 km.

Density $\rho_0 = 1.225\ \text{kg.m}^{-3}$ (6.6)

Molecular Mass (Dry Air) $m_a = 28.96442\ \text{kg.kmol}^{-1}$ (6.7)

Thermal Conductivity $k_0 = 2.5343 \times 10^{-2}\ \text{W.m}^{-1}.\text{K}^{-1}$ (6.8)

Dynamic Viscosity $\mu_0 = 1.7894 \times 10^{-5}\ \text{N.s.m}^{-2}$ (6.9)

Kinematic Viscosity $\nu_0 = 1.4607 \times 10^{-5}\ \text{m}^2.\text{s}^{-1}$ (6.10)

Universal Gas Constant $R_0 = 8314.32\ \text{J.kmol}^{-1}.\text{K}^{-1}$ (6.11)

Sutherland's Constants $S = 110.4\ \text{K}$ (6.12)

$\beta_s = 1.458 \times 10^{-6}\ \text{N.s.m}^{-2}.\text{K}^{-0.5}$ (6.13)

Specific Heat Capacity $C_p = 1005\ \text{J.kg}^{-1}.\text{K}^{-1}$ (6.14)

The parameter given in (6.14), namely C_P, is the specific heat capacity of a gas at constant pressure. An equivalent parameter, C_V, is defined at constant volume. The ratio of these parameters for dry air is defined as follows:

$$\gamma = C_p/C_v = 1.4 \tag{6.15}$$

The concentration of atmospheric gases can be approximated as given in Table 6.1. Bearing in mind that summary data of this type can vary in precision, the end result is a mean molecular mass of dry air that is usually quoted between 28.96 and 28.97 kg/kmol. In this case, the numbers conveniently add up to the relevant ISA constant although the detail is unimportant. In addition to the gas components included in this calculation, note that there are smaller concentrations of other gases (such as helium and methane) and, recognising that the ISA is based on dry air, the presence of water vapour is ignored.

As stated earlier, a standard atmosphere treats the air mass as a single entity that comprises a homogeneous mixture of constituent gases. For practical purposes, the same mixture exists up to 80 km and accordingly the value of m_a is constant. Thus the value of the gas constant $R = R_0/m_a$ applies to the whole atmosphere model:

$$R = \frac{R_0}{m_a} = \frac{8314.32}{28.96442} = 287.05287\ \text{J.Kg}^{-1}.\text{K}^{-1} \tag{6.16}$$

For the reasons discussed in Section 6.1, the atmospheric composition changes above 80 km. Although not of direct relevance to aviation, it is informative to see the impact of that change. This is defined in the US Standard Atmosphere and is depicted in Figure 6.2.

The speed of sound (a) in a gas is defined as follows:

$$a = \sqrt{\gamma R T} \tag{6.17}$$

where R is the gas constant, T is the gas temperature and γ is the ratio of specific heat capacities C_P and C_V. Note that the specific heat capacity is the amount of energy required in order to

Table 6.1 Approximate composition of dry air.

Gas	Molecular Mass	Fraction	Mass Fraction
N_2	28.013	78.083%	21.87339
O_2	31.999	20.946%	6.70251
Ar	39.948	0.933%	0.37272
CO_2	44.010	0.035%	0.01540
Ne	20.180	0.002%	0.00040
Molecular Mass (Dry Air)			**28.96442**

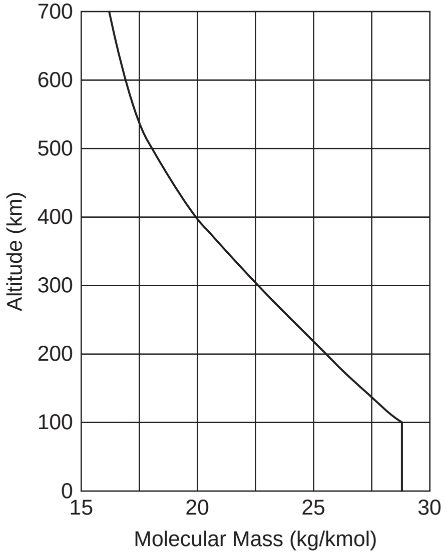

Figure 6.2 **Variation of molecular mass with altitude.**

effect a 1 K temperature rise per kilogram. In this case, this refers to a gas and the subscript ‘P’ denotes gas at constant pressure and the subscript ‘V’ denotes gas at constant volume. The ratio $\gamma = C_P/C_V$ has a value of 1.4 for dry air, as given in (6.15). Thus, the speed of sound in dry air can be derived as follows:

$$a = 20.046796\sqrt{T} \tag{6.18}$$

The relationship between pressure (p) and temperature (T) is:

$$p = \rho RT \tag{6.19}$$

where R is the gas constant and ρ is the gas density.

Thermal conductivity (k) is expressed as a function of temperature:

$$k = 2.648151 \times 10^{-3} \frac{T^{3/2}}{T + 245.4 \times 10^{(-12/T)}} \tag{6.20}$$

Dynamic viscosity (μ) is also expressed as a function of temperature:

$$\mu = \beta_s \frac{T^{3/2}}{T + S} \tag{6.21}$$

where S and β_s are Sutherland’s empirical constants, given in (6.12) and (6.13).

Substituting numerical values, this becomes:

$$\mu = 1.458 \times 10^{-6} \frac{T^{3/2}}{T + 110.4} \tag{6.22}$$

From this, kinematic viscosity (ν) is derived as:

$$\nu = \frac{\mu}{\rho} \tag{6.23}$$

where ρ is the gas density.

As well as using actual values of pressure (p), temperature (T) and density (ρ), it is often appropriate to use nondimensional parameters that relate actual values to reference values. In the context of atmospheric models, this is established using the ISA constants that are given in (6.4) to (6.6). Thus, the definitions of relative pressure (δ), relative temperature (θ) and relative density (σ) are given by:

$$\delta = \frac{p}{p_0} \tag{6.24}$$

$$\theta = \frac{T}{T_0} \tag{6.25}$$

$$\sigma = \frac{\rho}{\rho_0} \tag{6.26}$$

Note that, applying these parameters, Equations (6.17) and (6.19) can be recast as follows:

$$a = a_0\sqrt{\theta} \tag{6.27}$$

$$\delta = \sigma\theta \tag{6.28}$$

6.4 Geopotential Altitude

6.4.1 *Standard Definition*

The variation of atmospheric properties with altitude is based on the principle of *hydrostatic equilibrium*, i.e. the air is stationary with respect to the Earth. This postulates a thin block of air with thickness dh and with unit horizontal area. Pressure p acts from below and pressure p + dp acting from above. The difference in these pressures is equivalent to the reduction in weight of the vertical column of air that results from subtracting that block of air. Thus hydrostatic equilibrium is established in the form:

$$p = p + dp + \rho g\, dh$$

This simplifies (very simply!) to the standard form:

$$dp = -\rho g\, dh \tag{6.29}$$

This is known as the *hydrostatic equation*.

The value of gravitational acceleration (g) varies with altitude. As defined in (4.24), this obeys the inverse square law:

$$g = \frac{GM}{r^2} \tag{6.30}$$

As conventionally defined, r is the radial distance from the centre of mass and GM is the product of the gravitational constant and the mass of the Earth. Recall that WGS84 assigns a numerical value of $GM = 3.986004418 \times 10^{14}\ m^3s^{-2}$, as stated in (4.25). Assuming a mean radius of $r_0 = 6356766\ m$ (as given in (6.1)), this gives a familiar-looking nominal value for gravitational acceleration of about $9.86 m.s^{-1}$. This is greater than the corresponding ISA constant, which will be discussed later.

In order to avoid dealing with variations in gravity with altitude, (6.26) can be rewritten in an equivalent form, namely:

$$dp = -\rho g\ dh = -\rho g_0\ dH \tag{6.31}$$

Now the hydrostatic equation is based on the value of gravitational acceleration at sea-level and, as such, the altitude scale must be adjusted. The equivalence is expressed as:

$$dH = \frac{g}{g_0} dh \tag{6.32}$$

where h is the *geometric* altitude and H is the so-called *geopotential* altitude. At sea-level, (6.27) becomes:

$$g_0 = \frac{GM}{r_0^2} \tag{6.33}$$

At altitudes above sea-level, recognising that $r = r_0 + h$, the following integration can be performed:

$$H = \int_0^h \frac{g}{g_0} dh = \int_0^h \left(\frac{r_0}{r}\right) dh = \int_0^h \left(\frac{r_0}{r_0 + h}\right)^2 dh = -\left[\frac{r_0^2}{r_0 + h}\right]_0^h$$

The result is a simple relationship between geometric and geopotential altitudes:

$$H = \frac{r_0 h}{r_0 + h} \tag{6.34}$$

The corresponding inverse relationship is, as follows:

$$h = \frac{r_0 H}{r_0 - H} \tag{6.35}$$

6.4.2 *Generalised Definition*

The standard definition of geopotential altitude is based on ISA constants, which means that it is strictly valid at a single latitude north or south of the equator. A generalised definition needs to refer to geospatial and gravitational models from which those constants derive. In previous chapters, for example, this has been addressed via the WGS84 ellipsoid (cf. Section 6.2.1) and its associated gravity formula (cf. Section 4.2.9). In the current context (as defined in EDSU Data Items 77022b and 79018, and elsewhere), this is addressed via an 'International Ellipsoid of Reference' and a gravity formula that is called the Lambert Equation.

The parameters of the reference ellipsoid are, as follows:

$$\text{Earth Rotation Rate} \quad \omega = 7.29212 \times 10^{-5}\ rad.s^{-1} \tag{6.36}$$

$$\text{Equatorial Radius} \quad a = 6378178\,\text{m} \tag{6.37}$$

$$\text{Polar Radius} \quad c = 6356798\,\text{m} \tag{6.38}$$

The Lambert Equation specifies gravitational acceleration at mean sea level as a function of geodetic latitude[6] (φ):

$$g_0(\varphi) = 9.80616\left(1 - 0.0026373\cos 2\varphi + 0.0000059\cos^2 2\varphi\right) \tag{6.39}$$

Setting this quantity equal to ISA gravity as given in (6.2), this implies that ISA constants are valid for the nominal latitude of $\varphi = 45.5425°$.

The inverse square law for gravitation is stated in (6.30). The effect of altitude can be approximated by the following ratio:

$$g(\varphi) = g_0(\varphi)\left(\frac{r_0}{r_0 + h}\right)^2 \tag{6.40}$$

This specifies gravitational acceleration at altitude (h) above mean sea level as a function of geodetic latitude (φ). In fact, it measures altitude radially outwards from the centre of the Earth rather than orthogonal to the reference ellipsoid. A more accurate method relates to (2.76) and (2.77) but, for practical purposes, the eccentricity of the Earth is sufficiently small as to cause no great problem (as already discussed in Section 2.6.4).

Incorporating the centrifugal effects of Earth rotation (cf. Section 4.2.3), (6.39) is modified:

$$g(\varphi) = \left[g_0(\varphi) + \omega_0^2 r_0(\varphi)\cos^2\varphi\right]\left(\frac{r_0(\varphi)}{r_0(\varphi) + h}\right)^2 - \omega_0^2\left(r_0(\varphi) + h\right)\cos^2\varphi \tag{6.41}$$

However, rather than propagate the additional algebra, it is preferable to revert to the simpler structure of (6.40) and introduce a fictitious Earth radius (r_e). Thus,

$$g(\varphi) = g_0(\varphi)\left(\frac{r_e(\varphi)}{r_e(\varphi) + h}\right)^2 \tag{6.42}$$

where values of this fictitious radius are given by:

$$r_e(\varphi) = \frac{2g_0(\varphi)}{3.085462 \times 10^{-6} + 2.27 \times 10^{-9}\cos 2\varphi - 2 \times 10^{-12}\cos 4\varphi} \tag{6.43}$$

Note that, at the nominal latitude for ISA constants (discussed above), the numerical value is 6356766 m at $\varphi = 45.5425°$, which accounts for (6.1).

From consideration of hydrostratic equilibrium, the general version of (6.31) is:

$$dp = -\rho g(\varphi)\,dh = -\rho g_0(\varphi)\,dH$$

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[6] Note that geodetic latitude is denoted here by φ where, in Chapter 2, it was denoted by μ.

Accordingly, the equivalence with sea-level values (as originally stated in (6.32)) now becomes:

$$dH = \frac{g(\varphi)}{g_0(\varphi)} \left(\frac{g_0(\varphi)}{g_0} \right) dh \tag{6.44}$$

This integrates to give the generalised relationship between geometric and geopotential altitudes (cf. (6.34)), namely:

$$H = \frac{r_e(\varphi)\ h}{r_e(\varphi) + h} \left(\frac{g_0(\varphi)}{g_0} \right) \tag{6.45}$$

Remember that the gravitational acceleration at mean sea level $g_0(\varphi)$ is given by (6.39).

Recalling the discussion of geospatial geometry in Chapter 2, the radial distance from the centre of the Earth to any point on its surface is given by:

$$r_e(\varphi) = \sqrt{\frac{a^4 + c^4 \tan^2\varphi}{a^2 + c^2 \tan^2\varphi}} \tag{6.46}$$

or, equivalently:

$$r_e(\varphi) = \sqrt{\frac{a^4 \cos^2\varphi + c^4 \sin^2\varphi}{a^2 \cos^2\varphi + c^2 \sin^2\varphi}} \tag{6.47}$$

where the parameters where the parameters a and c are defined by (6.37) and (6.38) and where the notation from Chapter 2 has been modified to avoid a clash with ISA parameters.

6.5 Vertical Structure of the Atmosphere

Atmosphere models provide parametric data as functions of geopotential altitude. The vertical structure is based on a temperature profile that is appropriate for a given representation of the atmosphere. This takes the form of a multilayered model with a linear temperature variation within each layer. What distinguishes a particular model is the number layers, where the boundaries occur between layers and what temperature gradients are adopted within each layer. The general temperature-altitude relationship is defined as:

$$T = T_n + L_n(H - H_n) \tag{6.48}$$

This applies to the n-th layer with a base altitude H_n, a base temperature T_n (defined at H_n) and a linear gradient L_n (above H_n). Also, by implication, there is a base pressure p_n (defined at H_n).

The fundamental relationships governing pressure are given in (6.28) and (6.19):

$$dp = -\rho g_0\, dH$$

$$\rho = \frac{p}{RT}$$

These can be combined in order to give:

$$\frac{dp}{p} = -\frac{g_0}{R}\frac{1}{T}dH \tag{6.49}$$

Therefore,

$$\int \frac{dp}{p} = -\frac{g_0}{R}\int \frac{1}{T}dH \tag{6.50}$$

Applying the temperature profile from (6.48), there are two cases to be considered, namely $L_n = 0$ and $L_n \neq 0$.

When $L_n = 0$, the integral is trivial:

$$\log_e p - \log_e p_n = -\frac{g_0}{RT_n}\int_{H_n}^{H} dH = -\frac{g_0}{RT_n}(H - H_n)$$

$$\log_e\left(\frac{p}{p_n}\right) = -\frac{g_0}{RT_n}(H - H_n)$$

$$\frac{p}{p_n} = \exp\left(-\frac{g_0}{RT_n}(H - H_n)\right) \tag{6.51}$$

Alternatively, altitude can be expressed as a function of pressure ratio:

$$H - H_n = -\frac{RT_n}{g_0}\log_e\left(\frac{p}{p_n}\right) \tag{6.52}$$

When $L_n \neq 0$, the integral is nearly as trivial. From (6.33), it is seen that:

$$dT = L_n dH \tag{6.52}$$

Thus,

$$\log_e p - \log_e p_n = -\frac{g_0}{RL_n}\int_{T_n}^{T}\frac{dT}{T} = -\frac{g_0}{RL_n}(\log_e T - \log_e T_n)$$

$$\log_e\left(\frac{p}{p_n}\right) = -\frac{g_0}{RL_n}\log_e\left(\frac{T}{T_n}\right)$$

$$\frac{p}{p_n} = \left(\frac{T}{T_n}\right)^{-g_0/RL_{ni}} \tag{6.53}$$

Again, as an alternative, altitude can be expressed as a function of pressure ratio:

$$H - H_n = \frac{T - T_n}{L_n} = \frac{T_n}{L_n}\left(\frac{T}{T_n} - 1\right)$$

$$H - H_n = \frac{T_n}{L_n}\left[\left(\frac{p}{p_n}\right)^{-RL_n/g_0} - 1\right] \tag{6.54}$$

Using these expressions, the ISA vertical structure can be developed, using the values of H_n, T_n and L_n given in Table 6.2. The associated variations in pressure and temperature are shown in Figure 6.3. The underlying data for that figure was computed using Algorithm 6.1.

Table 6.2 Data for the International Standard Atmosphere.

n	H_n	T_n	L_n	p_n	g_0/RL_n	g_0/RT_n
0	0	288.15	-6.5×10^{-3}	1.013250×10^5	−5.255880	n/a
1	11000	216.65	0	2.263204×10^4	n/a	1.576885×10^{-4}
2	20000	216.65	1.0×10^{-3}	5.474879×10^3	34.16322	n/a
3	32000	228.65	2.8×10^{-3}	8.680160×10^2	12.20115	n/a
4	47000	270.65	0	1.109058×10^2	n/a	1.262266×10^{-4}
5	51000	270.65	-2.8×10^{-3}	6.693853×10^1	−12.20115	n/a
6	71000	214.65	-2.0×10^{-3}	3.956392	−17.08161	n/a
7	80000	196.65	n/a	8.862722×10^{-1}	n/a	n/a

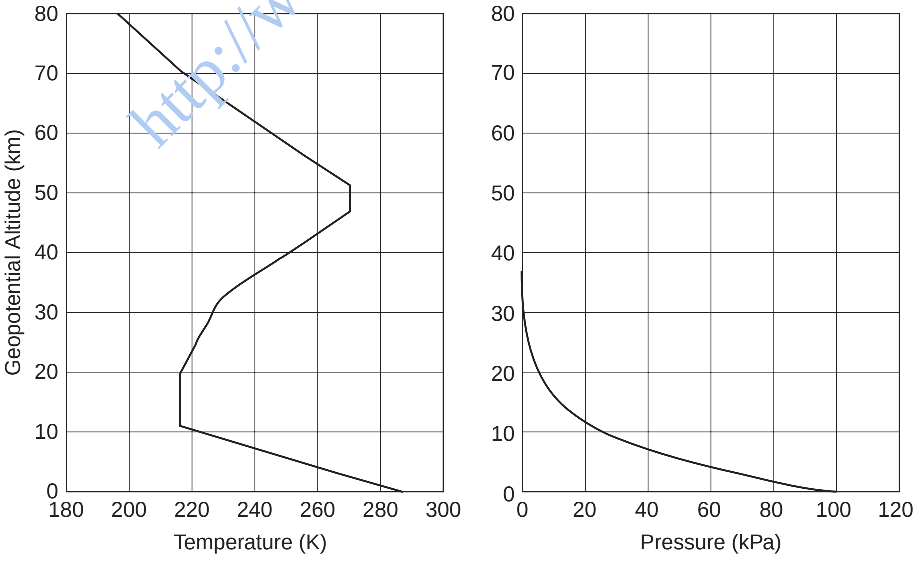

Figure 6.3 ISA temperature and pressure profiles.

Table 6.3 Pressure envelopes based on atmospheric measurement.

H (km)	High Pressure (mbar)					Low Pressure (mbar)				
	Max	1%	5%	10%	20%	Min	1%	5%	10%	20%
0	1084	–	–	–	–	870	–	–	–	–
1	930	920	918	917	916	842	847	856	861	868
2	821	817	816	815	814	736	742	748	752	757
4	643	642	641	640	639	548	550	558	565	569
6	501	499	497	496	495	406	408	413	418	422
8	385	384	383	382	381	296	299	303	306	309
10	294	293	292	291	290	215	218	221	223	225
12	226	226	224	223	222	154	157	158	160	162
14	168	167	166	165	164	111	111	113	115	117
16	123	123	122	121	120	79	79	80	82	84
18	88	88	87	86	85	56	56	57	58	59
20	65	65	64	63	62	40	41	42	43	45
22	45	45	44	44	43	28	29	30	31	32
24	35	34	33	32	32	20	21	24	27	28
26	26	25	25	24	24	14	15	18	20	21
28	20	19	19	18	18	10	11	13	15	16
30	15	15	14	14	13	7	9	10	11	12

Table 6.4 Temperature envelopes based on atmospheric measurement.

H (km)	High Temperature (°C)					Low Temperature (°C)				
	Max	1%	5%	10%	20%	Min	1%	5%	10%	20%
0	58	49	46	45	–	−68	−61	−57	−54	−51
1	41	40	39	38	34	−54	−53	−51	−50	−49
2	32	30	29	28	27	−47	−41	−36	−34	−31
4	19	17	14	13	12	−53	−48	−46	−44	−40
6	8	6	4	3	0	−61	−57	−56	−54	−51
8	−4	−5	−6	−9	−11	−68	−66	−65	−64	−61
10	−13	−13	−17	−19	−20	−75	−74	−72	−70	−65
12	−22	−22	−24	−30	−31	−80	−73	−72	−70	−67
14	−30	−30	−35	−36	−40	−77	−75	−74	−73	−70
16	−35	−37	−39	−39	−40	−87	−86	−84	−83	−82
18	−35	−37	−38	−39	−40	−88	−86	−84	−82	−81
20	−31	−32	−38	−39	−40	−87	−86	−83	−81	−71
22	−29	−30	−38	−38	−39	−85	−84	−83	−82	−82
24	−33	−33	−38	−39	−39	−86	−85	−83	−82	−81
26	−27	−28	−37	−37	−38	−84	−84	−83	−81	−80
28	−22	−23	−28	−33	−36	−84	−83	−81	−79	−78
30	−17	−18	−23	−28	−33	−85	−83	−81	−79	−76

Stacking layers of the atmospheric model together, the pressure and temperature at any altitude can be obtained as follows:

$$\frac{T}{T_0} = \left(\frac{T}{T_n}\right)\left(\frac{T_n}{T_{n-1}}\right)\left(\frac{T_{n-1}}{T_{n-2}}\right)\cdots\left(\frac{T_2}{T_1}\right)\left(\frac{T_1}{T_0}\right) \tag{6.55}$$

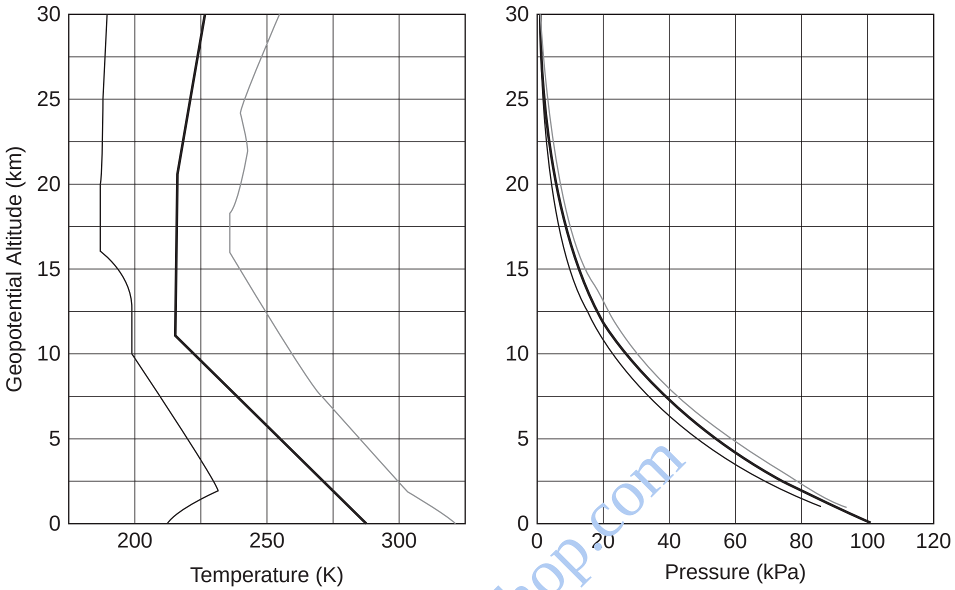

Figure 6.4 ISA profiles with 99 % likelihood envelopes.

$$\frac{P}{P_0} = \left(\frac{P}{P_n}\right)\left(\frac{P_n}{P_{n-1}}\right)\left(\frac{P_{n-1}}{P_{n-2}}\right)\cdots\left(\frac{P_2}{P_1}\right)\left(\frac{P_1}{P_0}\right) \tag{6.56}$$

This information can be set in an appropriate context by comparing it with data presented in MIL-HDBK-310. This provides a wide range of environmental data intended to guide the development of military products. Amongst other items, there are pressure and temperature envelopes based on measurement, as given in Tables 6.3 and 6.4. These do not represent coherent profiles but are simply maximum and minimum values at each altitude, together with related values for frequencies of occurrence of 1 %, 5 %, 10 % and 20 %. This is illustrated in Figure 6.4, where the ISA pressure and temperature profiles are contained within envelopes with 99 % likelihood of occurrence (i.e. between the 1 % values at the high and low extremes).

```
function (pressure,temperature) = InternationalStandardAtmosphere
  (altitude)
H = (0,11,20,32,47,51,71,80)*1000 ;
L = (-6.5,0.0,1.0,2.8,0.0,-2.8,-2.0)/1000 ;

isa.p0 = 101325 ;
isa.T0 = 288.15 ;
isa.g0 = 9.80665 ;
isa.R = 287.05287 ;

altitude = sort(altitude);
altitude(altitude<min(H)|altitude>max(H)) = [];
```

Continued

```
(P,T) = calculateIsaProfile(isa,H,L);
n = findLayer(altitude,H);
(pressure,temperature) = calculateProfile(isa,altitude,n,H,L,P,T);

function (P,T) = calculateIsaProfile(isa,H,L)
P(1) = isa.p0;
T(1) = isa.T0;
for iLayer = 2:length(H)
 gradient = L(iLayer-1);
 dH = H(iLayer) - H(iLayer-1);
 T(iLayer) = T(iLayer-1) + dH*gradient;

 switch gradient
 case 0
 exponent = -isa.g0*dH/(isa.R*T(iLayer-1));
 P(iLayer) = P(iLayer-1)*exp(exponent);
 otherwise
 exponent = -isa.g0/(isa.R*gradient);
 ratio = T(iLayer)/T(iLayer-1);
 P(iLayer) = P(iLayer-1)*ratio^exponent;
 end
end

function (pressure,temperature)=calculateProfile(isa,altitude,
  n,H,L,P,T)
for iAltitude = 1:length(altitude)
 iLayer = n(iAltitude);
 gradient = L(iLayer);
 dH = altitude(iAltitude) - H(iLayer);
 temperature(iAltitude) = T(iLayer) + dH*gradient;

 switch gradient
 case 0
 exponent = -isa.g0*dH/(isa.R*T(iLayer));
 pressure(iAltitude)=P(iLayer)*exp(exponent);
 otherwise
 exponent = -isa.g0/(isa.R*gradient);
 ratio = temperature(iAltitude)/T(iLayer);
 pressure(iAltitude)=P(iLayer)*ratio^exponent;
 end
end

function n = findLayer(altitude,H)
for iAltitude = 1:length(altitude)
 nLayer = max(find(altitude(iAltitude)>=H));
 switch nLayer
 case length(H)
 n(iAltitude) = nLayer - 1 ;
 otherwise
 n(iAltitude) = nLayer ;
 end
end
```

Algorithm 6.1 'InternationalStandardAtmosphere'.

6.6 Pressure Altitude

By definition, the *pressure altitude* of the point in any atmosphere is the geopotential altitude in the ISA that gives the same pressure. The ISA relationship between pressure and geopotential altitude is used as the calibration law for altimeters, which means that an altimeter will display pressure altitude if it is set to display zero when $p = p_0$, where p_0 is the ISA sea-level pressure defined in (6.4).

Combining (6.19) and (6.31), the relationship between pressure and geopotential altitude can be written as:

$$dp = -\frac{pg_0}{RT}\, dH \tag{6.57}$$

This can be rewritten specifically for ISA:

$$dp = -\frac{pg_0}{RT_{ISA}}\, dH_P \tag{6.58}$$

where pressure altitude (H_P) can be used because the temperature variation follows the ISA profile (as denoted by the subscript 'ISA'). Therefore the variation of geopotential altitude with pressure altitude is given by:

$$\frac{dH}{dH_P} = \frac{T}{T_{ISA}} \tag{6.59}$$

The ISA temperature profile is defined by (6.48) and is redefined here, as follows:

$$T_{ISA} = T_n + L_n(H_P - H_{Pn}) \tag{6.60}$$

where H_P is pressure altitude and H_{Pn} is the pressure altitude that corresponds with the base of the n-th layer in the atmosphere model. Any other temperature at any altitude can be produced simply by adding an increment (ΔT):

$$T = T_{ISA} + \Delta T \tag{6.61}$$

Thus, (6.59) becomes:

$$\frac{dH}{dH_P} = \frac{L_n(H_P - H_{Pn}) + T_n + \Delta T}{L_n(H_P - H_{Pn}) + T_n} \tag{6.62}$$

There are two cases to be considered, namely $L_n = 0$ and $L_n \neq 0$.

When $L_n = 0$, the integral is trivial:

$$\int_{H_n}^{H} dH = \frac{T_n + \Delta T}{T_n} \int_{H_{Pn}}^{H_P} dH_P$$

$$H - H_n = \left(1 + \frac{\Delta T}{T_n}\right)(H_P - H_{Pn})$$

$$= H_P - H_{Pn} + \left(\frac{\Delta T}{T_n}\right)(H_P - H_{Pn})$$

The relevant ISA relationship between geopotential altitude and pressure ratio was given by (6.52). By definition, this is equally valid for pressure altitude:

$$H_P - H_{Pn} = -\frac{RT_n}{g_0}\log_e\left(\frac{p}{p_n}\right) \tag{6.63}$$

Therefore, in the case when $L_n = 0$, the final expression that links geopotential altitude in any atmosphere with its equivalent pressure altitude is:

$$H - H_n = H_P - H_{Pn} - \Delta T\frac{R}{g_0}\log_e\left(\frac{p}{p_n}\right) \tag{6.64}$$

When $L_n \neq 0$, the integral is slightly more demanding:

$$\int_{H_n}^{H} dH = \int_{H_{Pn}}^{H_P}\left[1 + \frac{\Delta T}{L_n(H_P - H_{Pn}) + T_n}\right]dH_P$$

$$H - H_n = H_P - H_{Pn} + \frac{\Delta T}{L_n}\log_e\left[\frac{L_n(H_P - H_{Pn}) + T_n}{T_n}\right]$$

$$= H_P - H_{Pn} + \frac{\Delta T}{L_n}\log_e\left[1 + \frac{L_n}{T_n}(H_P - H_{Pn})\right]$$

The relevant ISA relationship between geopotential altitude and pressure ratio was given by (6.54). By definition, this is equally valid for pressure altitude:

$$H_P - H_{Pn} = \frac{T_n}{L_n}\left[\left(\frac{p}{p_n}\right)^{-RL_n/g_0} - 1\right] \tag{6.65}$$

A small amount of manipulation gives:

$$1 + \frac{L_n}{T_n}(H_P - H_{Pn}) = \left(\frac{p}{p_n}\right)^{-RL_n/g_0}$$

Therefore, in the case when $L_n \neq 0$, the final expression that links geopotential altitude in any atmosphere with its equivalent pressure altitude is:

$$H - H_n = H_P - H_{Pn} - \Delta T\frac{R}{g_0}\log_e\left(\frac{p}{p_n}\right) \tag{6.66}$$

Noting that (6.60) and (6.62) are identical, it is possible to write a single equation that defines the difference between geopotential altitude and pressure altitude:

$$H - H_P = H_n - H_{Pn} - \Delta T\frac{R}{g_0}\log_e\left(\frac{p}{p_n}\right) \tag{6.67}$$

This applies within the n-th layer of the atmosphere model. Lower layers are governed by equations of the form:

$$H_i - H_{Pi} = H_j - H_{Pj} - \Delta T \frac{R}{g_0} \log_e \left(\frac{p_i}{p_j} \right) \tag{6.68}$$

where $i = j+1$. Aggregating (6.67) and (6.68) across all layers in the atmosphere model, the following result is obtained:

$$H - H_P = H_0 - H_{P0} - \Delta T \frac{R}{g_0} \left[\log_e \left(\frac{p}{p_n} \right) + \log_e \left(\frac{p_n}{p_{n-1}} \right) + \ldots \log_e \left(\frac{p_2}{p_1} \right) + \log_e \left(\frac{p_1}{p_0} \right) \right]$$

$$H - H_P = H_0 - H_{P0} - \Delta T \frac{R}{g_0} \log_e \left(\frac{p}{p_0} \right) \tag{6.69}$$

Thus, the difference between *geopotential altitude* and *pressure altitude* at any altitude above sea-level is determined as the difference of these quantities at sea level decremented by a function of temperature increment (ΔT) (from (6.61)) and pressure ratio (δ) (from (6.24)). In appropriate notation, this becomes:

$$H - H_P = \Delta H_0 - \Delta T \frac{R}{g_0} \log_e \delta \tag{6.70}$$

where ΔH_0 is the geopotential altitude that corresponds with zero pressure altitude. By convention, sea-level pressure is standard in all reference atmospheres and so $H = H_0 = H_{P0} = 0$, i.e. $\Delta H_0 = 0$. Using the constants defined in Section 6.1, (6.70) reduces to the following practical formula:

$$H - H_P = \Delta H_0 - 29.271247 \Delta T \ln \delta \tag{6.71}$$

6.7 Reference Atmospheres

As discussed, a standard atmosphere model is constructed as a global average of atmospheric parameters over a full year. Variations due to regional, diurnal and seasonal factors are removed, as well as those due to latitude. Thus, a single model is available for performance analysis and instrument calibration. As has been seen already, the International Standard Atmosphere is valid for a geodetic latitude of 45.5425°.

Nonstandard *reference atmospheres* provide a wider context for aircraft design and operation, by characterising typical profiles that are associated with climatic zones and with seasonal variations are different latitudes. Further information can be found in ESDU Data Items 78008c and 78012b. In addition, a wide range of climatic and atmospheric data is presented in MIL-HDBK-310.

For current purposes it is appropriate to summarise the temperature profiles of four commonly quoted atmospheres in relation to the ISA profile, namely (1) Tropical Maximum, (2) Temperate and Arctic Maximum, (3) Tropical and Temperate Minimum and (4) Arctic Minimum. This is given in the form of a graphical sketch in Figure 6.5, with the underlying data contained in Table 6.5. Note that these data are calibrated against pressure altitude and only extend up to 20 km. Attention is drawn to the different number of layers in each

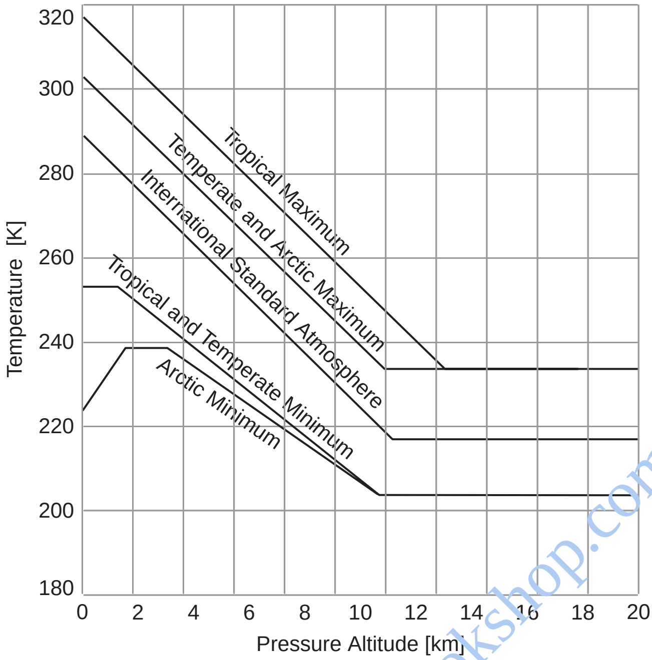

Figure 6.5 Temperature profiles for reference atmospheres.

Table 6.5 Data for reference atmospheres.

Atmosphere	n	H_{Pn}	T_n	L_n
International Standard Atmosphere (ISA)	0	0	288.15	−0.0065
	1	11000	216.65	0
	2	20000	216.65	n/a
Tropical Maximum	0	0	318.15	−0.0065
	1	13077	233.15	0
	2	20000	233.15	n/a
Temperate and Arctic Maximum	0	0	303.15	−0.0065
	1	10769	233.15	0
	2	20000	233.15	n/a
Tropical and Temperate Minimum	0	0	253.15	0
	1	1219	253.15	−0.0052917
	2	10668	203.15	0
	3	20000	203.15	n/a
Arctic Minimum	0	0	223.15	0.0098425
	1	1524	238.15	0
	2	3048	238.15	−0.0045932
	3	10668	203.15	0
	4	20000	203.15	n/a

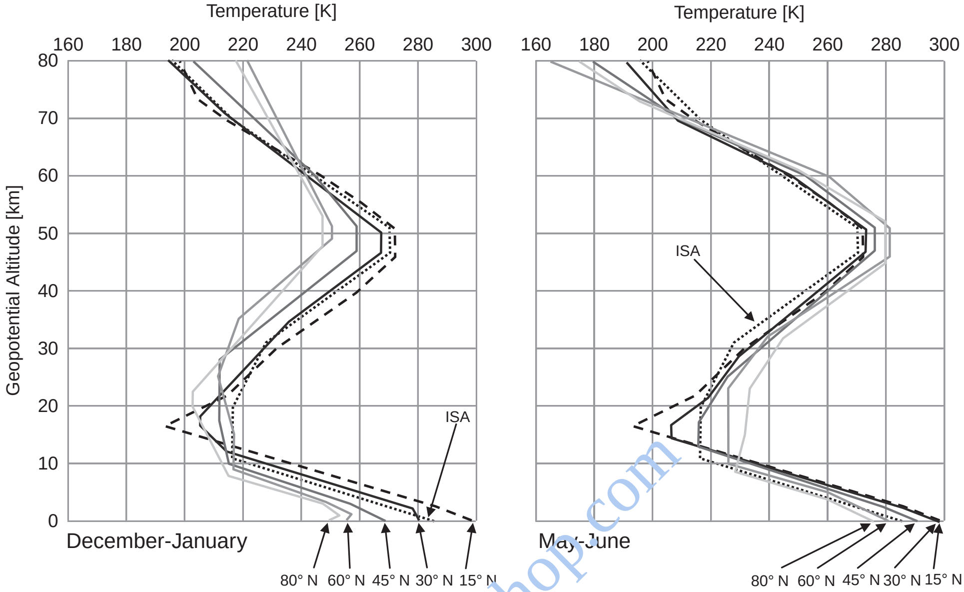

Figure 6.6 Mean temperature profiles at various latitudes.

atmosphere model, with specific emphasis on the low temperature near ground level in the Arctic Minimum climate.

6.8 Seasonal Variation

In order to illustrate the extent of seasonal variation, further information is sketched graphically in Figure 6.6. This shows the difference between midwinter climate (averaged over December and January) and midsummer climate (averaged over June and July) for a selection of northerly latitudes. For purposes of comparison, the ISA profile is included together with the mean annual profile for latitude 15 °N. Note that these data are calibrated against geopotential altitude and extend up to 80 km. Attention is drawn to the high number of layers in these atmosphere model, as opposed to the reference atmospheres giving global or climatic averages.

6.9 Climatic Regions

Standard atmospheres and reference atmospheres have the virtue of offering a small number of well-characterised *average* environments that can be used for engineering analysis and calibration. In order to set the context for this type of modelling, this section will give a very brief introduction to the climatic classification that defines the regional surface environments on the Earth. Two schemes will be considered, namely MIL-HDBK-310 and Köppen-Geiger.

6.9.1 *MIL-HDBK-310 Classification*

MIL-HDBK-310 is intended to give guidance on a wide of range of environmental factors that are relevant to military product design. In terms of surface environment, it defines five regional types of climate. Four of these pertain to land areas and are based on temperature and humidity, as these are important factors for equipment design. The four types of climatic region are designated Basic, Hot, Cold and Severe Cold.

The *Basic Region* is defined by 1 % values[7] of temperature of −31.7 °C and 43.3 °C. It relates to mid-latitudes, which would normally by thought of as having temperate or moderate climates, as well as humid tropics, which are warm throughout the whole year (without extremes of very high temperature). This is divided into five sub-regions with distinctive diurnal cycles, namely:

- Basic/Hot
- Basic/Cold
- Basic/Constant High Humidity
- Basic/Variable High Humidity
- Basic/Cold-Wet

Most attention is focused on the Basic Region because this type of climate applies is 'roughly coincident with the more densely populated, industrialised, and agriculturally productive areas of the world; therefore, most of the land areas with the highest probability of combat operations are within its limits' (MIL-HDBK-310, Section 6.2.1).

The Hot Region has a 1 % value of high temperature at 49 °C in its hottest places. This is subdivided into Hot/Dry and Hot/Humid diurnal cycles, which notionally coincide with deserts and hot tropics, respectively. The Cold Region has a 1 % value of low temperature at −45.6 °C in its coldest places. This is usually associated with high latitudes and high mountains. The Severe Cold Region has a 20 % value of low temperature at −51 °C and, effectively, represents polar climate.

Diurnal variations of temperature (in °C) and relative humidity (in percent) are summarised in Tables 6.6 and 6.7, respectively, for each type of climate region, with subdivisions as appropriate. The time-scale is given in hours, relating to local solar time. A diurnal cycle is not defined for the Severe Cold region because temperature and humidity remain almost constant.

It is clear that this information addresses a different class of problem from the atmosphere models discussed previously. This only applies to the surface environment and explicitly includes water vapour and daily fluctuations. In terms of aircraft performance, this would be appropriate for take-off and landing trials in a range of conditions (with particular emphasis on extreme conditions). Also, it would be appropriate for considering thermal performance of systems (especially engines) that are operating from start-up under different ambient or soak conditions.

6.9.2 *Köppen-Geiger Classification*

The Köppen classification is the most widely used framework for classifying climatic regions. It is based on the work of Wladimir Köppen from 1900 onwards, which was

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[7] A '1 % value' relates to a frequency of occurrence of 1 %. In other words, values that are less severe will be expected to occur on 99 % of occasions.

Table 6.6 Diurnal temperature cycles for MIL-HDBK-310 climatic regions.

Time Hrs	Basic region					Hot region		Cold region
	Hot	Cold	Constant high humidity	Variable high humidity	Cold-wet	Dry	Humid	
01	33	−31	24	27	−3	35	30	−46
02	32	−32	24	26	−3	34	31	−46
03	32	−32	24	26	−4	34	31	−46
04	31	−32	24	26	−4	33	31	−46
05	30	−32	24	26	−4	33	31	−46
06	30	−32	24	26	−4	32	32	−46
07	31	−30	24	27	−3	33	34	−45
08	34	−28	24	29	−2	35	36	−44
09	37	−26	24	31	−1	38	37	−43
10	39	−24	24	32	0	41	38	−41
11	41	−22	24	33	1	43	39	−39
12	42	−21	24	34	1	44	40	−37
13	43	−21	24	34	2	47	41	−37
14	43	−21	24	35	2	48	41	−37
15	43	−21	24	35	2	48	41	−37
16	43	−22	24	34	1	49	41	−38
17	43	−24	24	33	1	48	39	−39
18	42	−25	24	32	0	48	37	−39
19	40	−26	24	31	−1	46	36	−41
20	38	−27	24	29	−2	42	34	−42
21	36	−28	24	28	−2	41	33	−43
22	35	−29	24	28	−3	39	32	−44
23	34	−30	24	27	−3	38	32	−44
24	33	−31	24	27	−3	37	31	−45

Table 6.7 Diurnal humidity cycles for MIL-HDBK-310 climatic regions.

Time Hrs	Basic region					Hot region		Cold region
	Hot	Cold	Constant high humidity	Variable high humidity	Cold-wet	Dry	Humid	
01	36	N	100	100	100	6	88	N
02	38	E	100	100	100	7	88	E
03	41	A	100	100	100	7	88	A
04	44	R	100	100	100	8	88	R
05	44		100	100	100	8	88	
06	44	S	100	100	100	8	88	S
07	41	A	98	94	100	8	83	A
08	34	T	97	88	100	6	78	T
09	29	U	95	82	100	6	73	U
10	24	R	95	80	100	5	70	R
11	21	A	95	77	100	4	66	A
12	18	T	95	75	95	4	63	T

Table 6.7 (Continued).

Time Hrs	Basic region					Hot region		Cold region
	Hot	Cold	Constant high humidity	Variable high humidity	Cold-wet	Dry	Humid	
13	16	I	95	75	95	3	60	I
14	15	O	95	74	95	3	60	O
15	14	N	95	74	95	3	60	N
16	14		95	77	95	3	60	
17	14		95	79	100	3	64	
18	14		95	82	100	3	69	
19	17		97	86	100	3	74	
20	20		98	91	100	4	79	
21	22		100	95	100	5	85	
22	25		100	97	100	6	86	
23	28		100	98	100	6	87	
24	33		100	100	100	6	88	

probably motivated by patterns and delineations in vegetation. It is intended to reflect long-term mean climatic conditions and thus it gives an appropriate context for this book. Note that, for present purposes, it is an aside. The detail of this framework is less important than the recognition that a formal framework exists and is under continual development. This is of interest given the debate on climate change, especially with respect to the impact of aviation. It is also of interest when trying to make sense of the MIL-HDBK-310 climatic regions.

The good overview of climatic regions (and the associated issues of data and methodology) is presented by Peel *et al.* (2007), who describe their work on an updated world map for the Köppen-Geiger classification. It has the additional benefit of high-quality colour maps that allow the climatic distribution to be assimilated readily. The classification scheme (i.e. symbols and criteria) is defined in Table 6.8.

In Table 6.8 the following nomenclature is applied:

MAP = mean annual precipitation
MAT = mean annual temperature
T_{hot} = temperature of the hottest month
T_{cold} = temperature of the coldest month
T_{mon10} = number of months in which temperature is above 10 °C
P_{dry} = precipitation in driest month
P_{sdry} = precipitation in driest month of summer
P_{wdry} = precipitation in driest month of winter
P_{swet} = precipitation in wettest month in summer
P_{wwet} = precipitation in wettest month in winter
$P_{threshold}$ = 2xMAT (if 70% of MAP occurs in winter)
or 2xMAT+28 (if 70% of MAP occurs in summer)
or 2xMAT+14 (otherwise).

Note that Summer (winter) is defined as the warmer (cooler) six-month period of ONDJFM and AMJJAS.

Table 6.8 Köppen climate symbols and criteria.

Symbols			Description	Criteria
1st	2nd	3rd		
A			**Tropical**	$T_{cold} \geq 18$
	f		- Rainforest	$P_{dry} \geq 60$
	m		- Monsoon	Not(Af) & $P_{dry} \geq 100\text{-MAP}/25$
	w		- Savannah	Not(Af) & $P_{dry} < 100\text{-MAP}/25$
B			**Arid**	$MAP < 10xPthreshold$
	W		- Desert	$MAP < 5xPthreshold$
	S		- Steppe	$MAP \geq 5xPthreshold$
		h	-- Hot	$MAT \geq 18$
		k	-- Cold	$MAT < 18$
C			**Temperate**	$T_{hot} > 10$ & $0 < T_{cold} < 180$
	s		- Dry Summer	$P_{sdry} < 40$ & $P_{sdry} < P_{wwet}/3$
	w		- Dry Winter	$P_{wdry} < P_{swet}/10$
	f		- Without Dry Season	Not(Cs or Cw)
		a	-- Hot Summer	$T_{hot} \geq 22$
		b	-- Warm Summer	Not(a) & $T_{mon10} \geq 4$
		c	-- Cold Summer	Not(a or b) & $1 \leq T_{mon10} < 4$
D			**Cold**	$T_{hot} > 10$ & $T_{cold} \leq 0$
	s		- Dry Summer	$P_{sdry} < 40$ & $P_{sdry} < P_{wwet}/3$
	w		- Dry Winter	$P_{wdry} < P_{swet}/10$
	f		- Without Dry Season	Not(Ds or Dw)
		a	-- Hot Summer	$T_{hot} \geq 22$
		b	-- Warm Summer	Not(a) & $T_{mon10} \geq 4$
		c	-- Cold Summer	Not(a, b or d)
		d	-- Very Cold Winter	Not(a or b) & $T_{cold} < -38$
E			**Polar**	$T_{hot} < 10$
	T		- Tundra	$T_{hot} > 0$
	F		- Frost	$T_{hot} \leq 0$

As can be seen in the applicable criteria, this framework is based on averages of temperature and precipitation. The five major climate types can be summarised in everyday terms, as follows[8]:

A: Moist tropical climate with permanently high temperatures and very high rainfall
B: Dry climate with little rainfall and large diurnal variations in temperature
C: Humid mid-latitude climate in which land/water differences are significant and which have warm, dry summers and cool, wet winters
D: Continental climate in the interior of very large land areas, with modest rainfall and large seasonal variations in temperature
E: Very cold climate in which ice and tundra are permanent

The percentage distribution of major climate types for the world and for continents is presented in Table 6.9, as discussed by Peel *et al.* (2007). Here, the boundary between Asia and Europe is

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[8] www.blueplanetbiomes.org/climate.htm.

Table 6.9 Distribution of climate types.

Climate Type	A	B	C	D	E
Description	**Tropical**	**Arid**	**Temperate**	**Cold**	**Polar**
World	19.0%	30.2%	13.4%	24.6%	12.8%
Africa	31.0%	57.2%	11.8%		
Asia	16.3%	23.9%	12.3%	43.8%	3.8%
N America	5.9%	15.3%	13.4%	54.5%	11.0%
S America	60.1%	15.0%	24.1%		0.8%
Europe		36.3%	17.0%	44.4%	2.3%
Australia	8.3%	77.8%	13.9%		

Table 6.10 Proposed equivalence of climate types.

MIL-HDBK-310 climate type	Köppen-Geiger climate type	Description
HOT	BWh	Arid-Desert-Hot
	Cwa	Temperate-Dry Winter-Hot Summer
COLD	Dsc	Cold-Dry Summer-Cold Summer
	Dwc	Cold-Dry Winter-Cold Summer
	Dfc	Cold-Dry without Season-Cold Summer
	Dsd	Cold-Dry Summer-Very Cold Winter
VERY COLD	Dwd	Cold-Dry Winter-Very Cold Winter
	Dfd	Cold-Dry without Season-Very Cold Winter
BASIC	Remaining 22 types	*Refer to Table* 6.8

drawn nominally as a north–south line through the Ural Mountains down to the Arabian Sea. Also, North America is taken to include Central America and the West Indies.

It is observed that there are thirty possible climate types available in the full classification. The most extensive is BWh (Arid-Desert-Hot), covering 14.2 % of the land surface, followed by Aw (Tropical-Savannah), covering 11.5 %. At the other extreme, Cwc (Temperate-Dry Winter-Cold Summer) occurs over just 0.002 % of the land surface and Csc (Temperate-Dry Summer-Cold Summer) does not occur at all.

It is possible to identify a general mapping of climate types between Köppen-Geiger and MIL-HDBK-310, accepting the process is necessarily imprecise because different criteria are applied in each case. (Recall that the former classification combines temperature and precipitation while the latter combines temperature and humidity.) This is proposed in Table 6.10.

6.10 Air Density

The relationship between pressure (p) and temperature (T) has already been given in (6.19), with dependency on gas constant (R) and air density (ρ). Thus, air density can be written as:

$$\rho = \frac{p}{RT} \tag{6.72}$$

This is valid in all cases, for any atmosphere model. Following on from the discussion of pressure altitude in Section 6.5, this can be rewritten explicitly for ISA conditions, as follows:

$$\rho_{ISA} = \frac{p_{ISA}}{RT_{ISA}} \tag{6.73}$$

For any other atmosphere, the air density can be calibrated against pressure altitude, thereby allowing direct comparison with the ISA. Recalling (6.61), it is known that temperature variation from ISA conditions is established by introducing an increment (ΔT):

$$T = T_{ISA} + \Delta T$$

Therefore, (6.72) becomes:

$$\rho = \frac{p}{R(T_{ISA} + \Delta T)} \tag{6.74}$$

Picking two points with the same pressure altitude, such that $p = p_{ISA}$, it is clear that:

$$\rho = \frac{RT_{ISA}}{R(T_{ISA} + \Delta T)} \rho_{ISA}$$

$$\rho = \frac{1}{1 + \frac{\Delta T}{T_{ISA}}} \rho_{ISA} \tag{6.75}$$

For any other atmosphere, the air density at any point can be compared with the ISA via pressure altitude.

6.11 Water Vapour

Standard atmospheres and reference atmospheres all assume dry air, i.e. air without any water vapour. Environmental specifications and guidance documents, such as MIL-HDBK-310, most definitely do include water vapour. This raises the need to be able to account for the effect of water vapour in atmospheric calculations.

6.11.1 *Gas Constant*

The gas constant (R) of a homogeneous mixture of gases is derived as:

$$R = \frac{\sum R_i M_i}{\sum M_i} \tag{6.76}$$

where R_i is the gas constant for a particular component of the gas mixture and M_i is the mass of that component. The gas law (6.19) is applied to each component as:

$$p_i = \rho_i R_i T \tag{6.77}$$

or, alternatively,

$$p_i = \frac{M_i}{V} R_i T \tag{6.78}$$

This relates pressure (p_i) and temperature (T) for a mass of gas (M_i) occupying a fixed volume (V). In this context, the pressure of a particular component of a gas mixture is called the *partial pressure*.

Knowing that the gas constant is defined as

$$R_i = \frac{R_0}{m_i} \tag{6.79}$$

where R_0 is the *universal* gas constant and m_i is the molecular mass, the partial pressure can be rewritten as:

$$p_i = \frac{R_0 T}{V}\left(\frac{M_i}{m_i}\right) = \frac{R_0 T}{V} n_i \tag{6.80}$$

where n_i is the number of moles of gas. This demonstrates Avogadro's Law, which states that equal volumes of any gas contain equal numbers of atoms or molecules, under the assumption of equal temperature.

For a mixture of dry air and water vapour, this hybrid gas constant is derived in the following way:

$$R = \frac{R_a M_a + R_v M_v}{M_a + M_v} \tag{6.81}$$

where subscript 'a' denotes dry air and 'v' denotes water vapour.

The molecular masses of dry air and water vapour are:

$$m_a = 28.96442 \text{ kg.kmol}^{-1} \text{ (as defined in (6.7))} \tag{6.82}$$

$$m_v = 18.01594 \text{ kg.kmol}^{-1} \tag{6.83}$$

Recall that the universal gas constant is:

$$R_0 = 8314.32 \text{ J.kmol}^{-1}.\text{K}^{-1} \text{ (as defined in(6.11))}$$

Thus, the associated gas constants (from (6.79)) are given by:

$$R_a = 3287.05287 \text{ J.kg}^{-1}.\text{K}^{-1} \text{ (as derived in (6.16))} \tag{6.84}$$

$$R_v = 461.49798 \text{ J.kg}^{-1}.\text{K}^{-1} \tag{6.85}$$

It is often preferable to express (6.81) in terms of the partial pressures of dry air and water vapour. Given that partial pressure is defined by (6.78), the ratio of partial pressures in this case is defined by:

$$\frac{p_v}{p_a} = \frac{M_v}{M_a}\left(\frac{R_v}{R_a}\right) \tag{6.86}$$

Applying (6.79), this becomes:

$$\frac{p_v}{p_a} = \frac{M_v}{M_a}\left(\frac{m_a}{m_v}\right) \tag{6.87}$$

Equivalently, the mixture ratio is:

$$\frac{M_v}{M_a} = \frac{p_v}{p_a}\left(\frac{m_v}{m_a}\right) \tag{6.88}$$

With these equivalences in mind, the gas constant for the air/vapour mixture (cf. (6.81)) is:

$$R = \frac{R_a M_a + R_v M_v}{M_a + M_v} = R_a \frac{1 + \frac{M_v}{M_a}\left(\frac{R_v}{R_a}\right)}{1 + \frac{M_v}{M_a}}$$

$$R = R_a \frac{1 + \frac{p_v}{p_a}}{1 + \frac{p_v}{p_a}\left(\frac{m_v}{m_a}\right)} \tag{6.89}$$

Rearranging this expression, recognising that the total pressure $p = p_a + p_v$, the gas constant can be written in an alternative and frequently quoted form:

$$R = R_a \frac{p_a + p_v}{p_a + p_v\left(\frac{m_v}{m_a}\right)} = R_a \frac{p}{p - p_v + p_v\left(\frac{m_v}{m_a}\right)}$$

$$R = R_a \frac{p}{p - p_v\left(1 - \frac{m_v}{m_a}\right)} \tag{6.90}$$

Using the numerical values quoted in (6.82) and (6.83), this simplifies as follows:

$$R = \frac{287.05287\ p}{p - 0.377998\ p_v} \tag{6.91}$$

A method for calculating the partial pressure of water vapour is covered in the next section.

6.11.2 *Humidity*

Equations for the determination of humidity are given by Parish and Putnum (1977). This is based on vapour pressure (which is the partial pressure of water vapour in moist air) and saturated vapour pressure (at which two phases of water coexist in equilibrium). These pressures are *with respect to water (or ice)* when the air mass is over a plane of water (or ice) at the same temperature. In the case of saturated vapour pressure, any reduction in temperature will cause the formation of dew (or frost). Equivalently the *dew point* is the temperature at which the actual vapour pressure is equal to the saturated vapour pressure. If the dew point is at or below freezing, it is referred to as the frost point.

Absolute (or volumetric) humidity (AH) is the mass of water vapour per unit volume of air at a given temperature and pressure. With reference to the gas law given in (6.78), this is defined as follows:

$$\mathrm{AH} = \frac{p_v}{R_v T} \tag{6.92}$$

Relative humidity (RH) is defined as the ratio of vapour pressure (p_v) to saturated vapour pressure (p_s):

$$\mathrm{RH} = \frac{p_v}{p_s} \tag{6.93}$$

By recommendation, vapour pressure is evaluated with respect to water or ice (as appropriate) but saturated vapour pressure is always evaluated with respect to water.

The transformation from *water* to vapour is called *evaporation*; the transformation from *ice* to vapour is called *sublimation*. Both transformations require the addition of heat, or so-called latent heat (E), which is defined by:

$$\frac{dE}{dT} = C_{Pv} - C \tag{6.94}$$

where C_{Pv} is the specific heat capacity of water vapour at constant pressure and C is the specific heat capacity of water (or ice). As these parameters are almost constant for normal atmospheric pressures and temperatures, (6.94) becomes:

$$E = (C_P - C)(T - T_0) + E_0 \tag{6.95}$$

The Claperon-Clausius Equation relates pressure and temperature when two phases of matter are in equilibrium:

$$\left(\frac{1}{p_s}\right)\frac{dp_s}{dT} = \frac{E}{R_v T^2} \tag{6.96}$$

Thus,

$$\left(\frac{1}{p_s}\right)\frac{dp_s}{dT} = \frac{(C_{Pv} - C)\,(T - T_0) + E_0}{R_v T^2} = \frac{(C_{Pv} - C)}{R_v}\left(\frac{1}{T}\right) + \frac{E_0 - T_0(C_{Pv} - C)}{R_v}\left(\frac{1}{T}\right)^2$$

This integrates to give:

$$\log_e p_s = \frac{(C_{Pv} - C)}{R_v} \log_e T - \frac{E_0 - T_0(C_{Pv} - C)}{R_v}\left(\frac{1}{T}\right) + K \tag{6.97}$$

where K is the constant of integration. In general form, this can be rewritten as:

$$\log_e p_s = a \log_e T + b\left(\frac{1}{T}\right) + c \tag{6.98}$$

or equivalently:

$$p_s = T^a \exp\left[b\left(\frac{1}{T}\right) + c\right] \tag{6.99}$$

Now, by definition, the dew point is the temperature at which the actual vapour pressure is equal to the saturated vapour pressure. Therefore, (6.99) can be applied directly for vapour pressure (p_v) and drew point (D):

$$p_v = D^a \exp\left[b\left(\frac{1}{D}\right) + c\right] \tag{6.100}$$

Relevant constants in this context are:

Reference Temperature:	$T_0 = 273.15\,K$	(6.101)
Specific Heat Capacity:		
Ice at 0°C	$Ci = 2060\ J.kg^{-1}.K^{-1}$	(6.102)
Water at 0°C	$C_w = 4185\ J.kg^{-1}.k^{-1}$	(6.103)
Water Vapour	$C_{Pv} = 1911\ J.kg^{-1}.{-1}^{-1}$	(6.104)
Vapour Pressure:		
Ice at 0°C	$p_{I0} = 6.107$ mbar	(6.105)
Water at 0° C	$p_{w0} = 6.11$ mbar	(6.106)
Latent Heat:		
Sublimation	$E_{i0} = 2834000\ J.kg^{-1}$	(6.107)
Evaporation	$E_{w0} = 2500000\ J.kg^{-1}$	(6.108)

Note that pressure is measured in millibars.[9] Constants that appear in (6.98) and (6.99) need to be evaluated with respect to ice and water, using separate instances in (6.97):

$$\log_e p_{i0} = \frac{(C_{Pv} - C_i)}{R_v} \log_e T_0 - \frac{E_{i0} - T_0(C_{Pv} - C_i)}{R_v}\left(\frac{1}{T_0}\right) + K_i$$

$$\log_e p_{w0} = \frac{(C_{Pv} - C_w)}{R_v} \log_e T_0 - \frac{E_{w0} - T_0(C_{Pv} - C_w)}{R_v}\left(\frac{1}{T_0}\right) + K_w$$

---

[9] 1 bar = 100000 Pa (as distinct from 1 *atmosphere* = 101325 Pa).

These expressions relate to separate instances of (6.98):

$$\log_e p_s = a_i \log_e T + b_i\left(\frac{1}{T}\right) + c_i \tag{6.109}$$

$$\log_e p_s = a_w \log_e T + b_w\left(\frac{1}{T}\right) + c_w \tag{6.110}$$

where

$$\begin{aligned} a_i &= -0.322862 \\ b_i &= -6229.0616 \\ c_i &= 26.425244 \end{aligned}$$

and

$$\begin{aligned} a_w &= -4.927432 \\ b_w &= -6763.0699 \\ c_w &= 54.212470 \end{aligned}$$

In accordance with recommendation, saturated vapour is evaluated as:

$$p_s = T^{a_w} \exp\left[b_w\left(\frac{1}{T}\right) + c_w\right] \tag{6.111}$$

Actual vapour pressure is evaluated as:

$$p_v = D^{a_w} \exp\left[b_w\left(\frac{1}{D}\right) + c_w\right] \quad \text{when } D > 273.15 \text{ K} \tag{6.112}$$

$$p_v = D^{a_i} \exp\left[b_i\left(\frac{1}{D}\right) + c_i\right] \quad \text{when } D \le 273.15 \text{ K} \tag{6.113}$$

Note 6.1

For purposes of calculation, Equations (6.111) to (6.113) are not particularly well-conditioned in the sense that the first part of the evaluation (T^a) typically gives numbers of the order 10^{-13} while the second part typically gives numbers of the order 10^{12}. This can be resolved by using a normalised temperature, as defined in (6.25) and extended here to cover the dew point temperature:

$$\theta = \frac{T}{T_0} \quad \text{and} \quad \Delta = \frac{D}{T_0}$$

Also, there is often a preference for common logarithms in calculation rather than natural logarithms, and therefore terms of the form e^x (or $\exp(x)$) will have to be substituted by terms of the form 10^x.

Thus, the saturated vapour pressure (cf. (6.111)) is evaluated as:

$$p_s = \theta^{a_w} 10^{\left[b_w\left(\frac{1}{\theta}\right) + c_w\right]}$$

Actual vapour pressure (cf. (6.112) and (6.113)) is evaluated as:

$$p_v = \Delta^{a_w} 10^{\left[b_w\left(\frac{1}{\Delta}\right)+c_w\right]} \quad \text{when } \Delta > 1$$

$$p_v = \Delta^{a_i} 10^{\left[b_i\left(\frac{1}{\Delta}\right)+c_i\right]} \quad \text{when } \Delta \leq 1$$

In this context, the relevant constants become:

$$\begin{aligned} a_i &= -0.322862 \\ b_i &= -9.903888 \\ c_i &= 10.689717 \end{aligned}$$

and

$$\begin{aligned} a_w &= -4.927432 \\ b_w &= -10.752935 \\ c_w &= 11.538976 \end{aligned}$$

Note that these equations give vapour pressure measured in millibars, which means that the corresponding partial pressure for dry air (in the absence of any other component) is given by:

$$p_a = 1000 - p_v \tag{6.114}$$

Thus, in this case, the mixture ratio (6.88) can be expressed as:

$$\frac{M_v}{M_a} = \left(\frac{m_v}{m_a}\right)\frac{p_v}{1000 - p_v} = 0.622002\frac{p_v}{1000 - p_v} \tag{6.115}$$

To illustrate, Figure 6.7 shows the variation with temperature of (1) the saturated vapour pressure (in millibars) and (2) the vapour/air mixture ratio (in grams of vapour per kilogram of dry air). The variation of dew point (or frost point if it occurs at of below O°C) is shown in Figure 6.8.

The calculation of vapour pressure (for Figure 6.9) is undertaken using Algorithm 6.2. This code implements the set of equations specified in Note 6.1, via a generic computation with specialisation depending on whether the vapour pressure is with respect to water or ice. Temperature input is specified in degrees Celsius and all necessary conversion and normalisation is conducted internally. This algorithm transforms ambient temperature into saturated vapour pressure (recalling that this is always calculated with respect to water) and transforms the dew point temperature (if known) into actual vapour pressure.

Given that different equations apply to *water* and *ice*, an additional constant [d] is introduced in order to ensure that the switch-over occurs numerically at O °C. Using the original constants (a,b,c) to their given resolution, the switch actually occurs at 0.04 °C. Two iterations (using the residual error on each occasion) provided the necessary correction and, hence, the parameter ‘ice.d’ is composed of two numbers summed together. While this is an *ad hoc* procedure and as such might be open to criticism, it is always important (vitally important) that any transfer between different methods of calculation should occur at the intended transition, with as small a discontinuity as possible.

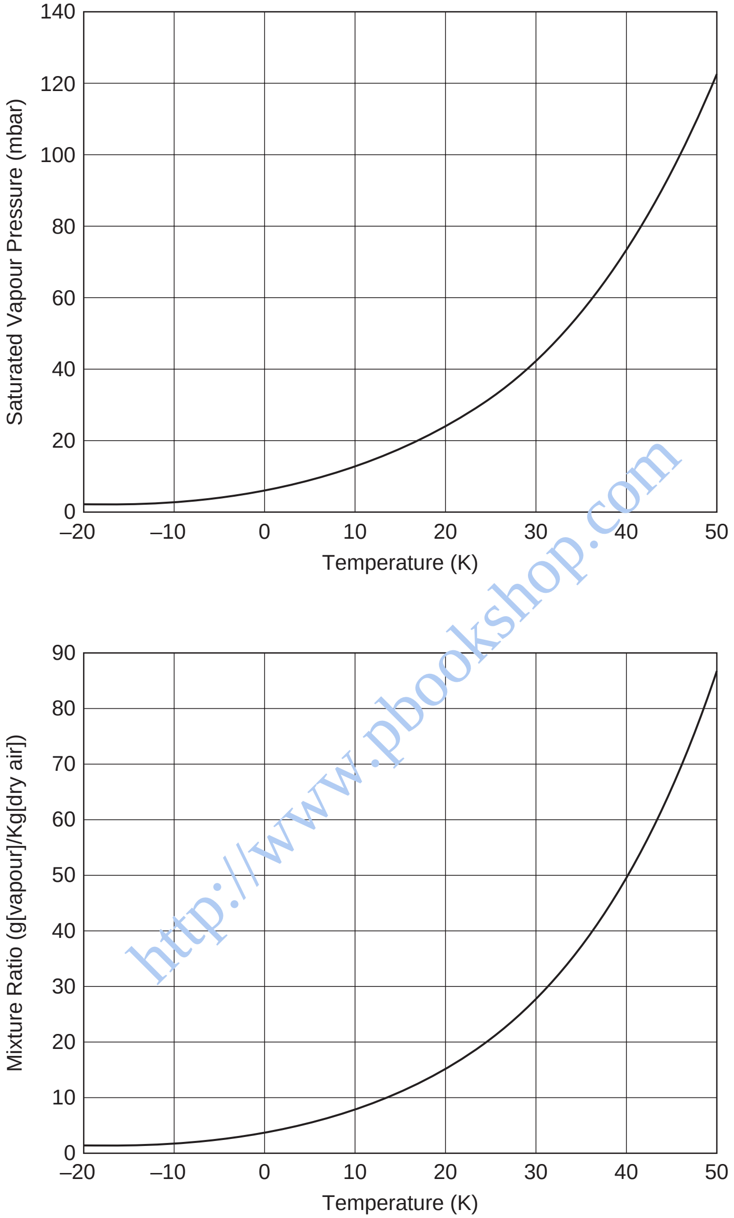


Figure 6.7 Vapour pressure and mixture ratio for saturated air.

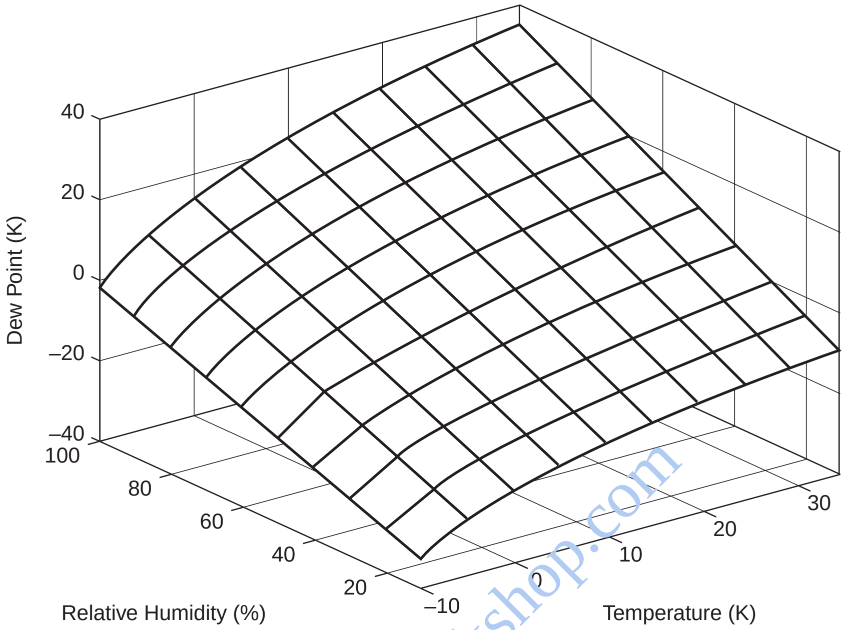


Figure 6.8 Dew point (frost point) variation vs ambient temperature and relative humidity.

```
function (p_value,p_deriv) = calculateVapourPressue(T,phase)

kelvin = @(celsius) celsius + 273.15 ;
normalise = @(T) T./273.15 ;

t = normalise(kelvin(T)) ;

ice.a = –0.322862 ;
ice.b = –2.903888 ;
ice.c = 10.689717 ;
ice.d = 0.00298185522617 + 2.664535259100376e–015 ;

water.a = –4.927432 ;
water.b = –10.752935 ;
water.c = 11.538976 ;
water.d = 0.0 ;

switch phase
 case 'water'
 constant = water ;
 case 'ice'
 constant = ice ;
 otherwise
 error('Phase must be ''water'' or ''ice''') ;
end
```

Continued

```
exponent = constant.c + constant.b./t ;
p_value = t.^constant.a .* 10.^exponent + constant.d ;

factor = constant.a - log(10)*constant.b./t ;
p_deriv = t^(constant.a-1) * 10.^exponent * factor ;
p_deriv = normalise(p_deriv) ;
```

Algorithm 6.2 'calculateVapourPressure'.

Environmental data often specifies ambient temperature and relative humidity. This requires a method for determining the dew point (or frost point). An example implementation is given in Algorithm 6.3. Current temperature is used to determine the saturated vapour pressure (via Algorithm 6.2) and the actual vapour pressure is determined from the relative humidity (as defined in (6.93)). This pressure is then used to find the dew point, based initially on the assumption that the vapour pressure is with respect to water; if the result is below 0 °C, the computation is repeated such that the vapour pressure is with respect to ice.

```
function D = calculateDewPoint(T,H)

fraction = @(percent) percent/100 ;
between = @(number,lower,upper) min(upper,max(lower,number)) ;

H = fraction(between(H,0,100)) ;

saturationPressure = calculateVapourPressure(T,'water') ;
vapourPressure = H * saturationPressure ;
D = calculateTemperature(vapourPressure,'water') ;
if isempty(D) || D<=0
 D = calculateTemperature(vapourPressure,'ice') ;
End
```

Algorithm 6.3 'calculateDewPoint'.

The determination of dew point is a temperature calculation that is implemented in Algorithm 6.4. This solves (6.112) and (6.113) (using the corresponding equations given in Note 6.1) by applying an iterative *gradient search* technique (refer to Note 6.2) because there is no convenient explicit formula for temperature as a function of pressure. The technique is highly effective and convergence should be rapid given the form of numerical relationship, as shown in Figure 6.7. The algorithm is simple but needs to be protected from circumstances that may prevent convergence, almost certainly involving very low vapour pressures. Thus, checks are made for zero gradient (to prevent division by zero) and for an unreasonably large number of iterations (to prevent an infinite loop). In situations where a result cannot be found, an empty result is returned. Thus, the computation always terminates in a controlled manner.

```
function T = calculateTemperature(vapourPressure,phase)

T = 100 ;

maxError = 1e-6 ;
nIteration = 0 ;
maxIterations = 50 ;
```

```
iterating = 1 ;
while iterating
 nIteration = nIteration + 1 ;
 (y,m) = calculateVapourPressure(T,phase) ;
 dy = y - vapourPressure ;

 if m==0 || nIteration>maxIterations
 T = [] ;
 iterating = 0 ;
 elseif abs(dy)<=maxError
 iterating = 0 ;
 else
 dT = dy/m ;
 T = T - dT ;
 end
end
```

Algorithm 6.4 'calculateTemperature'.

Note 6.2

The gradient search technique is one approach to solving nonlinear equations of the form $y = f(x)$. For a given value of y, several attempts are made to find a value of x that satisfies the equation. Because this is a numerical procedure, solutions are approximate except in special cases. Thus, the final value of x will be such that

$$|\ y - f(x)\ | < \varepsilon$$

i.e. the magnitude of the error between y and f(x) is less than a small number, ε.

With reference to the diagram, the initial guess is $x = x_0$. The initial error is

$$\Delta y_0 = f(x_0) - y$$

The initial gradient is determined as m_0 where

$$m_0 = \frac{\Delta y_0}{\Delta x_0}$$

The associated change in x is calculated as

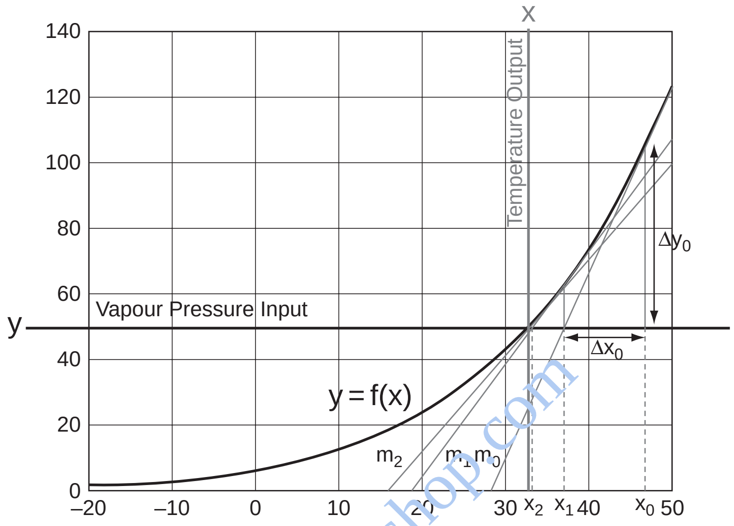
$$\Delta x_0 = \frac{\Delta y_0}{m_0}$$

Thus the next value of x is chosen to be:

$$x_1 = x_0 - \Delta x_0$$

Looking at the diagram, this steps closer the solution (where the line y = input intersects with the line y = f(x)). The process is repeated until convergence has been achieved. As can be seen, convergence is rapid in this particular application.

(continued)



6.12 Weather Systems

In contrast to 'average' atmospheric models, weather systems provide a time-varying environment for aircraft flight, both in terms of the dynamics of the air at any given location at any given time and in terms of the progress along a flight path between locations. So the requirement is to provide time-varying weather data covering a three-dimensional (or 3-D) space, thereby yielding a so-called four-dimensional (or 4-D) model. Specific data can then be interpolated along a given flight path, with time variation calibrated against path length.

In this book, the real interest lies with the types of weather information that are relevant to aviation and some of the considerations that relate to data availability. A readily accessible source of aviation weather information is provided by the Aviation Weather Center,[10] which is part of the NOAA[11] National Weather Service.[12] Data downloads are available via the Aviation Digital Data Service (ADDS).[13]

A number of example charts are presented in this section. Figure 6.9 shows wind speeds and streamlines at the surface, 12,000 ft, 24,000 ft and 36,000 ft. What is immediately striking in this case is that surface winds do not follow the trends of winds aloft and, also, that average wind speeds increase with altitude. Note the presence of winds in excess of

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[10] aviationweather.gov.

[11] National Oceanic and Atmospheric Administration.

[12] www.nws.noaa.gov.

[13] adds.aviationweather.gov.

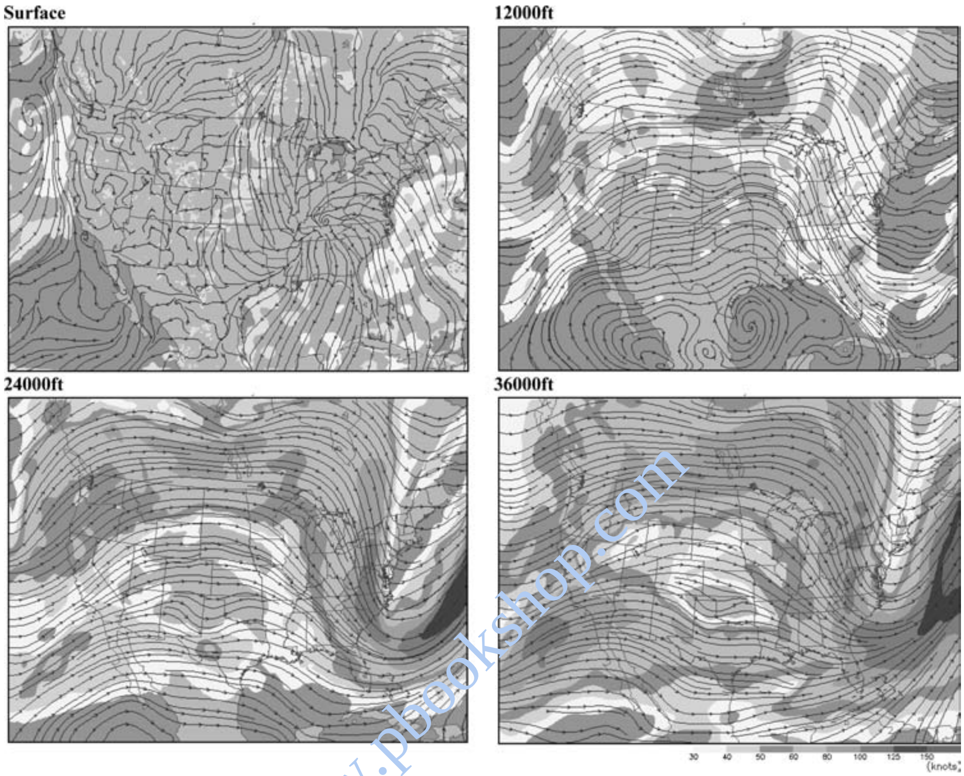

Figure 6.9 Example wind speed and streamline charts.

140 kts. Figure 6.10 shows the temperatures corresponding with those same altitudes. Note the distinction at surface level between land and sea temperatures. Above ground, given the wide application of the International Standard Atmosphere (ISA) (or, equivalently, the ICAO[14] Standard Atmosphere), it is convenient to quote temperatures with respect to the respective ISA datum values (i.e. −8.3 °C at 12,000 ft, −31.6 °C at 24,000 ft, −56.5 °C at 36,000 ft).

In this book, the real interest lies with the types of weather information that are relevant to aviation and some of the considerations that relate to data availability. A readily accessible source of aviation weather information is provided by the Aviation Weather Center[15], which is part of the NOAA[16] National Weather Service.[17] Data downloads are available via the Aviation Digital Data Service (ADDS).[18]

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[14] International Civil Aviation Organisation.

[15] aviationweather.gov.

[16] National Oceanic and Atmospheric Administration.

[17] www.nws.noaa.gov.

[18] adds.aviationweather.gov.

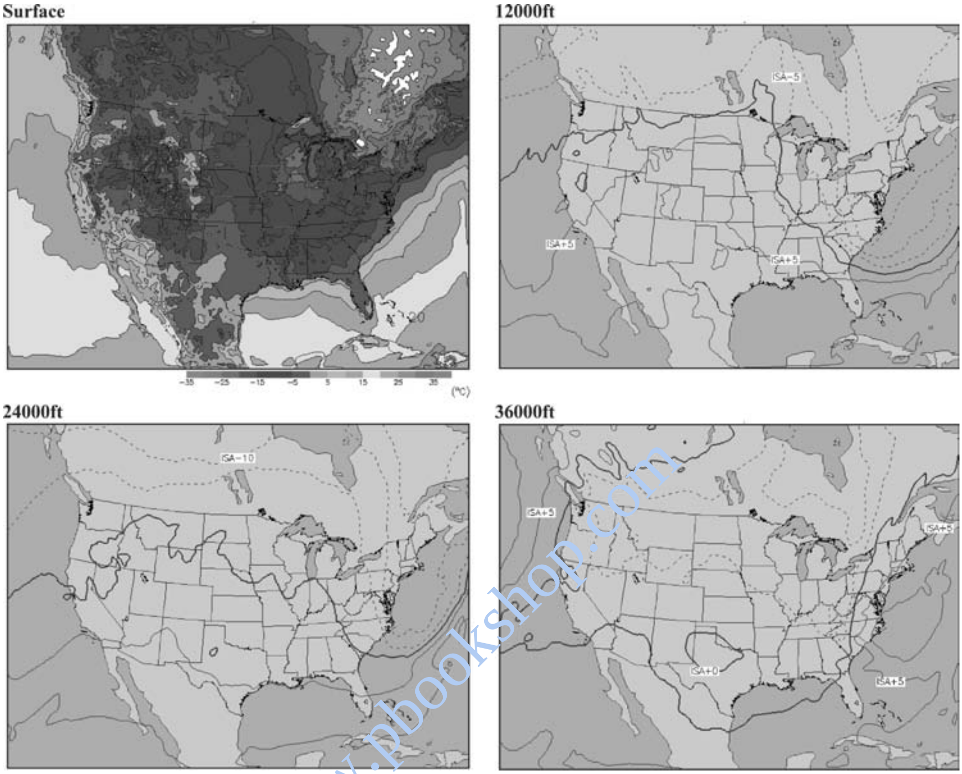

Figure 6.10 Example atmospheric temperature charts.

```
DATA BASED ON 030600Z
VALID 031200Z FOR USE 0800-1500Z. TEMPS NEG ABV 24000

FT 3000 6000 9000 12000 18000 24000 30000 34000 39000
ABR 2149 2438+12 2532+06 2527-02 2833-20 2940-32 293949 294458 284468
```

Figure 6.11 Typical wind/temperature text download format.

The ADDS offers text download for winds and temperatures in the form of a so-called NCEP Winds Aloft Forecast. A example of which is given in Figure 6.11. This is for low-level (up to 39,000 ft); a high-level forecast is available up to 53,000 ft. The header is decoded as follows:

```
DATA BASED ON 030600Z
```

Data based on computer forecasts generated for Day 3 of the month at 0600 UTC (cf. Section 3.3.3).

```
VALID 031200Z FOR USE 0800-1500Z. TEMPS NEG ABV 24000
```

Forecast is valid for Day 3 of the month at 1200 UTC and data can be used between 0800 and 1500 UTC. Temperatures are negative above 24,000 ft.

```
FT 3000 6000 9000 12000 18000 24000 30000 34000 39000
```

‘FT’ denotes the forecast location and the numbers denote altitudes.

The main body of data then follows. The forecast location is specified by its call sign and can be found by searching a database of NOAA weather stations.[19] The numerical data is given a number string for each altitude. For example, ‘2220 + 04’ means that the wind forecast is 220° at 20 kts and the temperature forecast is +4 °C. As stated in the header, temperatures are negative above 24,000 ft and so the number sign is not used for higher altitudes.

Wind direction is specified in 10° increments. Thus, when wind speed is less than 100 kt, wind direction is encoded as a number between 01 and 36 (in increments of 10°) and wind speed is encoded between 00 and 99. When wind speed exceeds this range, wind direction is encoded between 51 and 86 and wind speed is encoded between 00 and 99. Thus, in order to obtain actual values, subtract 50 from the first pair of digits and add 100 to the second pair of digits. For example, a data item ‘770249’ for 34,000 ft signifies a wind forecast of 270° at 102 kt and a temperature forecast of –49 °C.

Actual wind speeds are encoded between 5 kt and 199 kt; slower winds are treated as ‘zero’ and faster winds are treated as ‘199 knots or greater’. The code sequence becomes ‘9900’ (meaning winds light and variable) and, for example, ‘2599’ (meaning 250° at over 199 kt).

The example given in Figure 6.13 produces data for Aberdeen Regional Airport, Aberdeen, South Dakota. This is decoded to produce the results in Table 6.11. For reference, data on this particular forecast station is presented in Table 6.12. From a modelling perspective, this allows the data to be associated with an altitude scale projecting upwards from a geographical location.

Table 6.11 Wind/temperature forecast for Aberdeen Regional Airport, South Dakota (valid for 1200UTC, 3 January 2008).

Altitude (ft)	Wind Direction (deg)	Wind Speed (kt)	Temperature (°C)
3000	210	49	not defined
6000	240	38	12
9000	250	32	6
12000	250	27	−2
18000	280	33	20
24000	290	40	32
30000	290	39	49
34000	290	44	58
39000	280	44	68

---

[19] lwf.ncdc.noaa.gov/oa/climate/stationlocator.html.

Table 6.12 NOAA Weather Station Data for Aberdeen Regional Airport, SD.

Aberdeen Rgnl Airport
Aberdeen, SD, United States

Type	ASOS-NWS COOP WSO-N ASOS ABC NEXRAD
Call Sign/ICS	ABR/KABR
WBAN	14929
COOP ID	390020
Climate Division	SD-03 – Northeast
WMI ID	72659
In Service	01 Aug 1934 to Present
Elevation	395.3m (1297′) above s/l
Lat/Lon	45°27′N / 98°25′W
County	Brown

Data on surface pressure distribution is available from weather charts in the form of isobars with centres of High and Low pressure regions, together with weather fronts. The average surface pressure is nominally equal to the ISA constant given in (6.4), which is usually expressed as 1013 mbar in this context. The obvious practical issue is that an aircraft flying along a path at constant pressure altitude will actually be flying along a path of varying geometric (or geopotential) altitude. Equivalently, an altimeter would not display zero when at mean sea level.

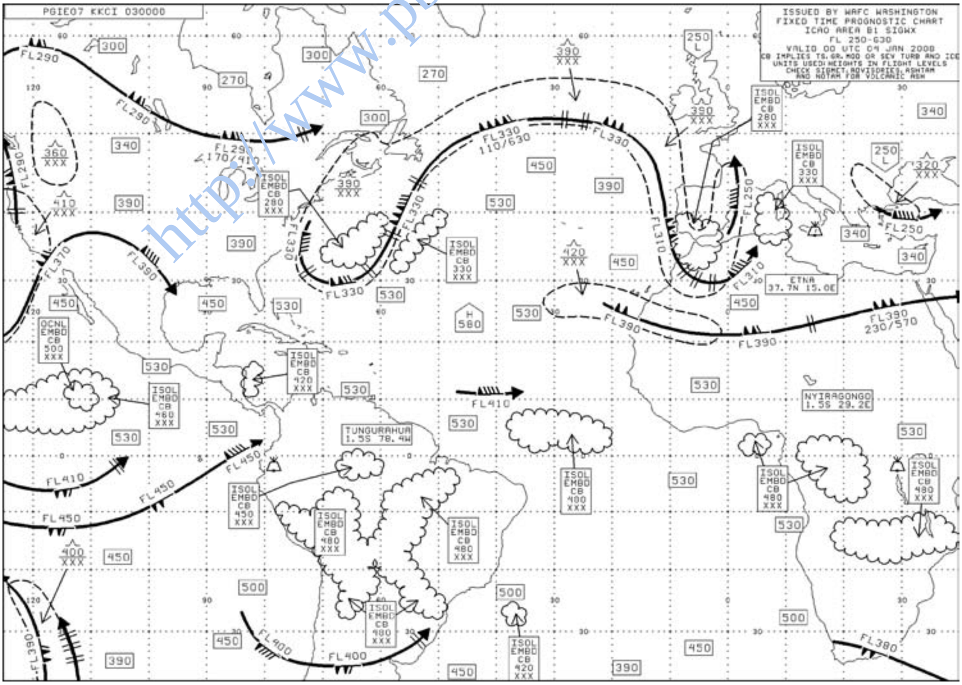

Figure 6.12 Example high-level significant weather chart.

The problem of atmospheric pressure variation is handled by allowing the datum pressure for the altimeter to be selected, depending on local or regional conditions. Two common settings are known as QNH and QFE. The QNH value is the pressure at mean sea level for a particular region and is required in order to establish height relative to topographical features identified on aeronautical charts. The QFE value is the pressure at a given field elevation and is required in order that the altimeter will read height above ground level during take-off and landing. Altimeters are set to the ISA datum above a threshold altitude (which is 3000 ft in the UK) and, in this setting, an altitude is expressed as a Flight Level (FL). Note that flight levels are resolved to 100 ft increments, such that 36,000 ft would be referred to as FL360.

Figure 6.12 shows a High-Level Significant Weather (SWH) chart. This type of information applies to flight levels between FL250 and FL630 above and is intended for the en-route sections of long-range flights. In addition, there are Medium-level Significant Weather (SWM) charts that cover the range FL100 to FL450. Collectively these are known as Significant Weather (SIGWX) charts and are issued by the World Area Forecast Centres (WAFCs) in London and Washington. The underlying data format is the Binary Universal

Table 6.13 BUFR features for SWH charts.

Feature	Designator
Jet-stream	JET
Clear Air Turbulence	TURB
Cloud	CLOUD
Tropopause Height	TROP
Tropical Cyclone and Sandstorms	STORM
Volcanic Eruption	VOLCANO
Radiation Event	RADIATION

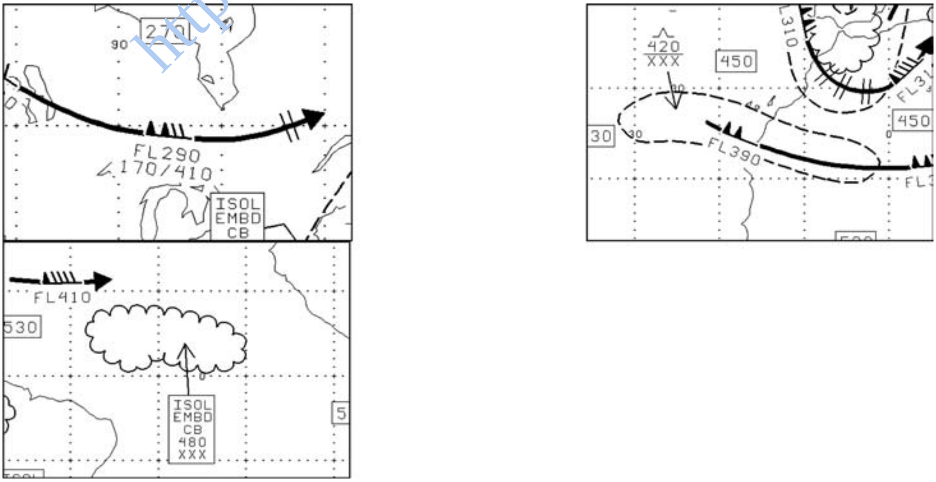

Figure 6.13 Main graphical features for SWH charts (JET,TURB,CLOUD).

Form for the Representation of meteorological data (BUFR). This is neutral data standard for SIGWX data that is defined in WMO[20] Manual 306 Part B. The features contained in SWH data are summarised in Table 6.13 and the main drawing features are shown in Figure 6.13.

A jet-stream is defined by a directed path, wind symbols and change bars. By ICAO definition this starts and stops at 80 kt. Wind symbols show direction via a line segment and speed via one or more fleches. Fleches are of three types, namely a triangle (denoting a 50 kt increment), a line (10 kt) and a short line (5 kt). The level of the maximum jet speed is shown, together wih the vertical depth (given by the lower and upper levels of the 80 kt isotachs). Thus, in the left pane of Figure 6.13, a 120 kt jet-stream is located at FL290, with 80 kt isotachs at FL170 and FL410. Change bars are orthogonal to the jet-stream, each denoting a 20 kt speed change.

Clear Air Turbulence (CAT) is represented by a dashed area, together with an index number or summary data. In the centre pane of Figure 6.13, the CAT area is the lozenge-shaped feature and the summary data (with an arrow pointing into the area) consists of a chevron (indicating moderate turbulence) together with a ceiling at FL420 and a base below FL250. Cloud is represented by a scalloped area, marked out by latitude and longitude, with an associated information box. In the right pane of Figure 6.13, the cloud area is annotated as ISOL/EMBED CB (isolated/embedded cumulo-nimbus) with a ceiling at FL480 and a base below FL250. Tropopause heights are shown by three-digit numbers enclosed within rectangular boxes, expressing height as a flight level.

A further set of descriptions, definitions and illustrations are available for SWM charts. More information is provided in a guidance document entitled 'Representing WAFS Significant Weather (SIGWX) Data in BUFR' Version 4.1,[21] issued by the WAFCs under joint patronage of the NOAA and UK Meteorological Office.

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[20] World Meteorological Organisation.

[21] Accessible via the Help page associated with aviationweather.gov/products/swh/.