4.3 **Upper wind conditions**

It is often observed that clouds at different levels move in different directions. The wind speeds at these levels may also differ markedly, although this is not so evident to the casual observer. The gradient of wind velocity with height is referred to as the (vertical) wind shear, and in the free air, above the friction level, the amount of shear depends upon the vertical temperature profile. This important relationship is illustrated in Figure 4.6. The diagram shows hypothetical contours of the 1000 and 500 mb pressure surfaces. As discussed in A.1 above, the thickness of the 1000 to 500 mb layer is proportional to its mean temperature: low thickness values correspond to cold air, high thickness values to warm air. This relationship is shown in Figure 4.1. The theoretical wind vector (VT) blowing parallel to the thickness lines, with a velocity proportional to their gradient, is termed the thermal wind. The geostrophic wind velocity at 500 mb (G500) is the vector sum of the 1000 mb geostrophic wind (G1000) and the thermal wind (VT), as shown in Figure 4.6.
The thermal wind component blows with cold air (low thickness) to the left in the northern hemisphere when viewed downwind; hence the poleward decrease of temperature in the troposphere is associated with a large westerly component in the upper winds. Furthermore, the zonal westerlies are strongest when the meridional temperature gradient is at a maximum (winter in the northern hemisphere).

The total result of the above influences is that in both hemispheres the mean upper geostrophic winds are dominantly westerly between the subtropical high- pressure cells (centred aloft at about 15° latitude) and the polar low-pressure centre aloft. Between the sub- tropical high-pressure cells and the equator the winds are easterly. The dominant westerly circulation reaches maximum speeds of 45 to 65 m s–1, which even increase to 135 m s–1 in winter. These maximum speeds are concentrated in a narrow band, often situated at about 30° latitude between 9000 and

15000 m, called the jet stream. Plate 14 shows bands of cirrus cloud that may have been related to jet-stream systems. The jet stream is essentially a fast-moving ribbon of air, connected with the zone of maximum slope, folding or fragmentation of the tropopause; this in turn coincides with the latitude of maximum poleward temperature gradient, or frontal zone, shown schematically in Figure 4.7. The thermal wind, as described above, is a major component of the jet stream, but the basic reason for the concentration of the meridional temperature gradient in a narrow zone (or zones) is dynamical. In essence, the temperature gradient becomes accentuated when the upper wind pattern is confluent.

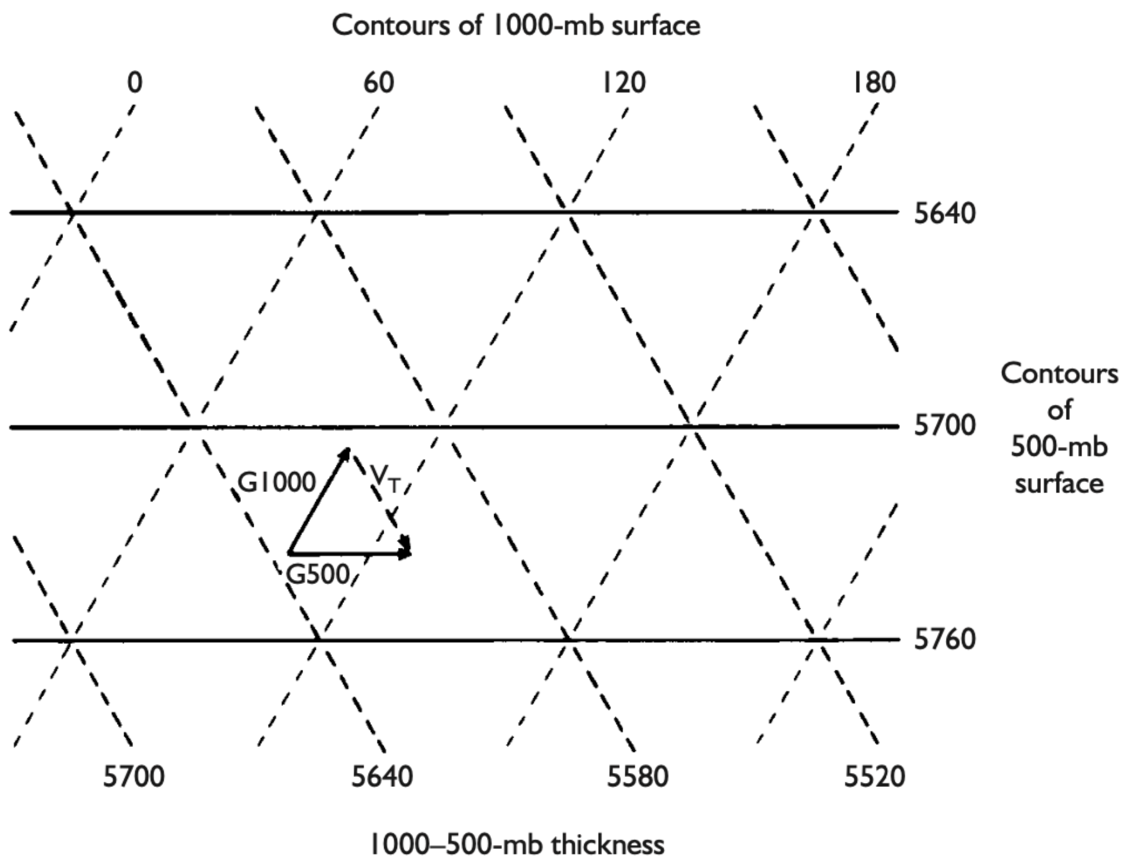

Figure 4.6 Schematic map of super- imposed contours of isobaric height and thickness of the 1000 to 500-mb layer (in metres). G_{1000} is the geostrophic velocity at 1000 mb, G_{500} that at 500 mb; V_T is the resultant 'thermal wind' blowing parallel to the thickness lines.

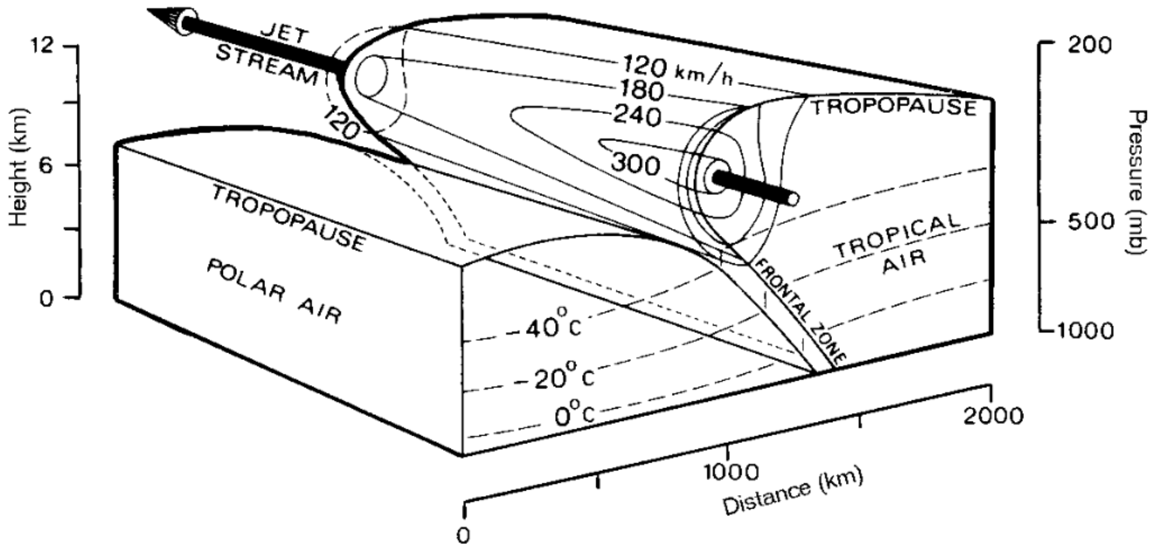

Figure 4.7 Structure of the mid-latitude frontal zone and associated jet stream showing generalized distribution of temperature, pressure and wind velocity. *Source*: After Riley and Spolton (1981)

Figure 4.8 shows a north–south cross-section with three westerly jet streams in the northern hemisphere. The more northerly ones, termed the polar front and arctic front jet streams, are associated with the steep temperature gradient where polar and tropical air and polar and arctic air, respectively, interact, but the subtropical jet stream is related to a temperature gradient confined to the upper troposphere. The polar front jet stream is very irregular in its longitudinal location and is commonly discontinuous (Plate 15), whereas the subtropical jet stream is much more persistent. For these reasons, the location of the mean jet stream in each hemisphere and season (Plate D) reflects primarily the position of the subtropical jet stream. The austral summer (DJF) map shows a strong zonal feature around 50°S, while the boreal summer jet is weaker and more discontinuous over Europe and North America. The winter maps (Plate D, [A] and [D]) show a pronounced double structure in the southern hemisphere from 60°E eastward to 120°W, a more limited analogue over the eastern and central North Atlantic Ocean (0 to 40°W). This double structure represents the subtropical and polar jets.

The synoptic pattern of jet stream occurrence may be complicated further in some sectors by the presence of additional frontal zones, each associated with a jet stream. This situation is common in winter over North America. Comparison of Figures 4.3, 4.4 and Plate D indicates that the main jet-stream cores are associated with the principal troughs of the

Rossby long waves. In summer, an easterly tropical jet stream forms in the upper troposphere over India and Africa due to regional reversal of the S–N temperature gradient. The relationships between upper tropospheric wind systems and surface weather and climate will be considered below.

In the southern hemisphere, the mean jet stream in winter is similar in strength to its northern hemisphere winter counterpart and it weakens less in summer, because the meridional temperature gradient between 30° and 50°S is reinforced by heating over the southern continents (Plate D).