Chapter 4. Planetary-scale motions in the atmosphere and ocean

4.1 A Variation of Pressure and Wind Velocity with Height

In this chapter, we examine global-scale motions in the atmosphere and their role in redistributing energy, momentum and moisture. As noted in Chapter 3 (p. 59), there are close links between the atmosphere and oceans with the latter making a major contribution to poleward energy transport. Thus, we also discuss ocean circulation and the coupling of the atmosphere–ocean system.

The atmosphere acts rather like a gigantic heat engine in which the temperature difference existing between the poles and the equator provides the energy supply needed to drive the planetary atmospheric and ocean circulation. The conversion of heat energy into kinetic energy to produce motion must involve rising and descending air, but vertical movements are generally less obvious than horizontal ones, which may cover vast areas and persist for periods of a few days to several months. We begin by examining the relationships between winds and pressure patterns in the troposphere and those at the surface.

Both pressure and wind characteristics change with height. Above the level of surface frictional effects (about 500 to 1000 m), the wind increases in speed and becomes more or less geostrophic. With further height increase, the reduction of air density leads to a general increase in wind speed (see Chapter 6A.1). At 45°N, a geostrophic wind of 14 m s–1 at 3 km is equivalent to one of 10 m s–1 at the surface for the same pressure gradient. There is also a seasonal variation in wind speeds aloft, these being much greater in the northern hemisphere during winter months, when the meridional temperature gradients are at a maximum. Such seasonal variation is absent in the southern hemisphere. In addition, the persistence of these gradients tends to cause the upper winds to be more constant in direction.

4.1.1 **The vertical variation of pressure systems**

The air pressure at the surface, or at any level in the atmosphere, depends on the weight of the overlying air column. In Chapter 2B, we noted that air pressure is proportional to air density and that density varies inversely with air temperature. Accordingly, increasing the temperature of an air column between the surface and, say, 3 km will reduce the air density and therefore lower the air pressure at the surface without affecting the pressure at 3 km altitude. Correspondingly, if we compare the heights of the 1000 and 700 mb pressure surfaces, warming of the air column will lower the height of the 1000 mb surface but will not affect the height of the 700 mb surface (i.e. the thickness of the 1000 to 700 mb layer increases).

The models of Figure 4.1 illustrate the relationships between surface and tropospheric pressure conditions. A low-pressure cell at sea-level with a cold core will intensify with elevation, whereas one with a warm core will tend to weaken and may be replaced by high pressure. A warm air column of relatively low density causes the pressure surfaces to bulge upward, and conversely a cold, more dense air column leads to downward contraction of the pressure surfaces. Thus, a surface high-pressure cell with a cold core (a cold anticyclone), such as the Siberian winter anticyclone, weakens with increasing elevation and is replaced by low pressure aloft. Cold anticyclones are shallow and rarely extend their influence above about 2500 m. By contrast, a surface high with a warm core (a warm anticyclone) intensifies with height (Figure 4.1D). This is characteristic of the large subtropical cells, which maintain their warmth through dynamic sub- sidence. The warm low (Figure 4.1C) and cold high (Figure 4.1B) ,whereas the other two types are produced primarily by dynamic processes. The high surface pressure in a warm anticyclone is linked hydrostatically with cold, relatively dense air in the lower stratosphere. Conversely, a cold depression (Figure 4.1A) is associated with a warm lower stratosphere.

Mid-latitude low-pressure cells have cold air in the rear, and hence the axis of low pressure slopes with height towards the colder air to the west. High-pressure cells slope towards the warmest air (Figure 4.2). Thus, northern hemisphere subtropical high-pressure cells are shifted 10 to 15° latitude southward at 3 km, and towards the west. Even so, this slope of the high- pressure axes is not constant through time.

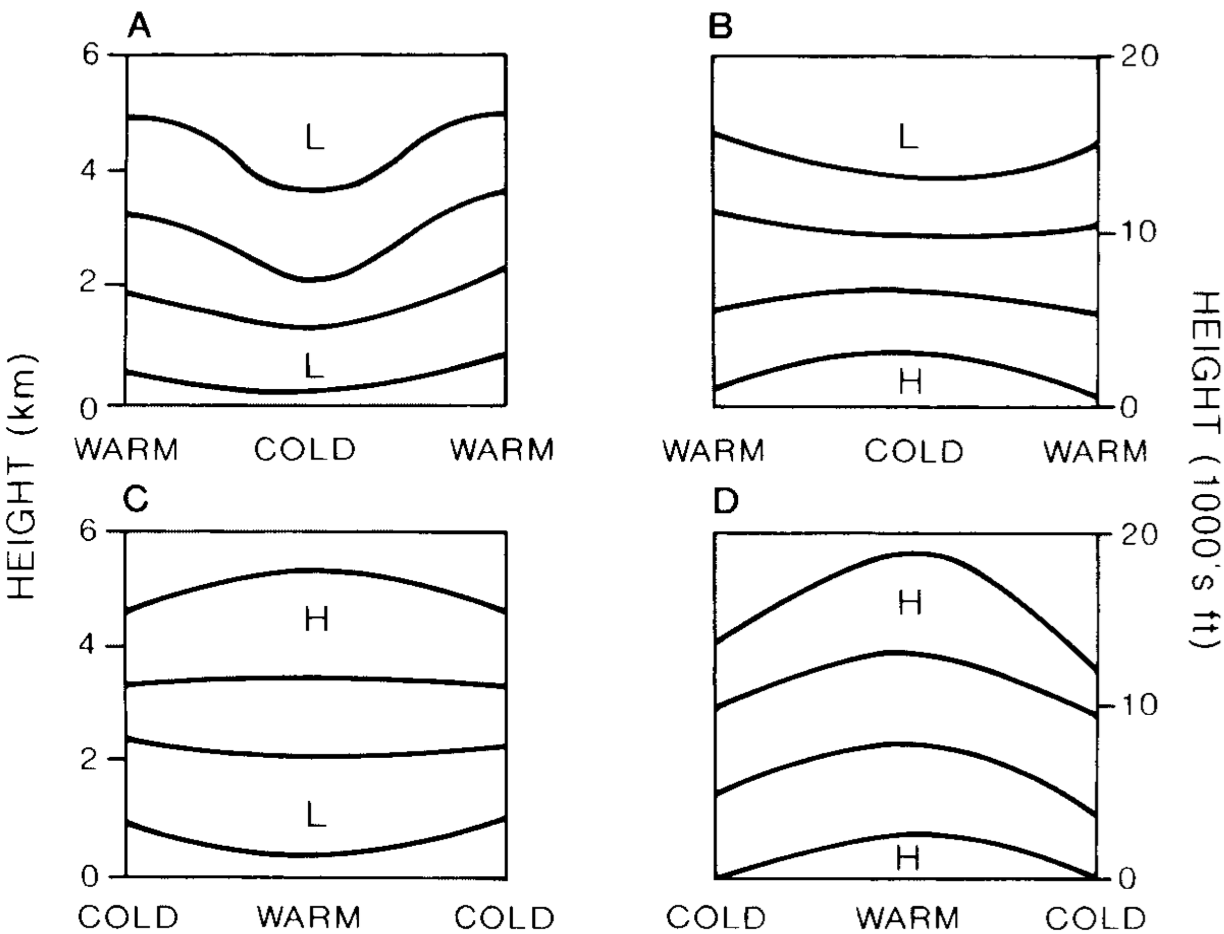

Figure 4.1 Models of the vertical pressure distribution in cold and warm air columns. (A) A surface low pressure intensifies aloft in a cold air column. (B) A surface high pressure weakens aloft and may become a low pressure in a cold air column. (C) A surface low pressure weakens aloft and may become a high pressure in a warm air column. (D) A surface high pressure intensifies aloft in a warm air column.

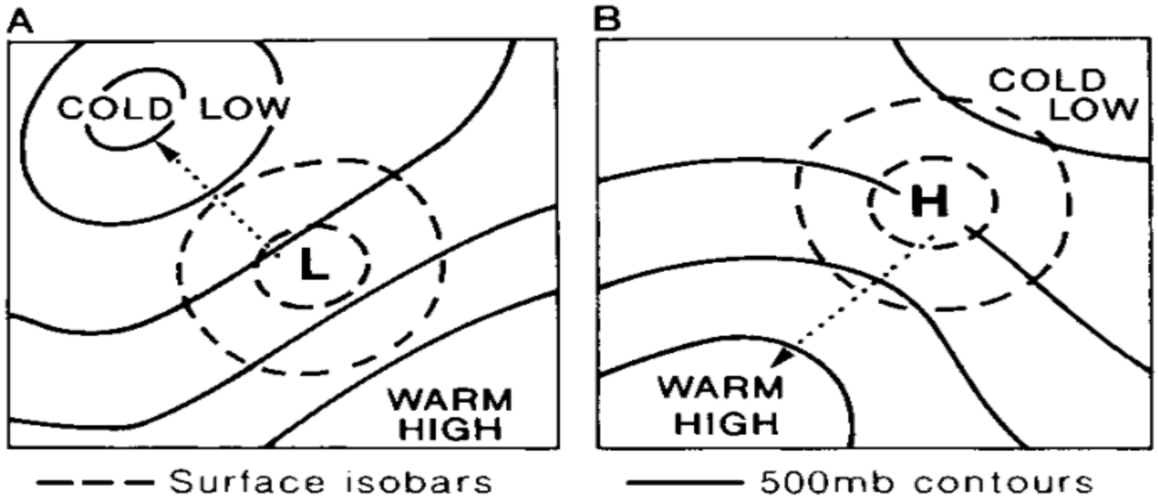

Figure4.2 The characteristic slope of the axes of low-and high- pressure cells with height in the northern hemisphere.