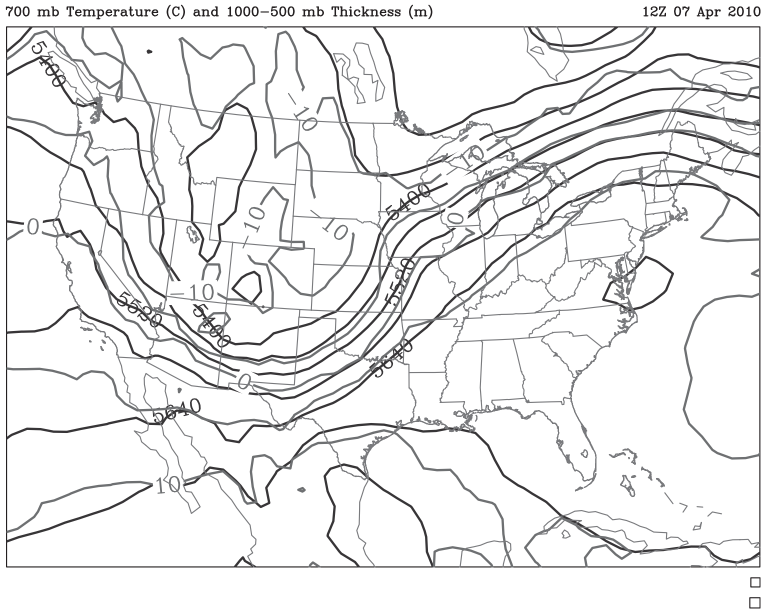

Fig. 2.2 Distributions of temperature at 700 mb (red) and thickness of the layer between 500 mb and 1000 mb (black) on April 7, 2010. See color plates section.

The bar denotes the average value of a quantity in a layer. This relationship is useful for weather forecasting, especially for the thickness of the layer between the 500-mb and 1000-mb surfaces, $\Delta Z = Z_{500-1000}$. A map of ΔZ is called a *thickness map*. The spacing between two adjacent contours of a thickness map is a visual indication of the spatial gradient of the mean temperature in the lower half of the troposphere. The smaller such spacing is, the stronger the thermal gradient would be. Figure 2.2 shows that the contours of the 700 mb mean temperature and those of the 500–1000 mb thickness on April 7, 2010 are indeed virtually parallel to one another.

(v) Sea level pressure differential over continent versus over oceans

It is well known observationally that the surface pressure over a large continent such as N. America or Asia is generally high in winter and low in summer. The reverse is true over the adjacent N. Pacific and N. Atlantic oceans. Before we can meaningfully compare the pressure distribution over continents with that over oceans, we first need to take into account the dependency of surface pressure on the elevation of the surface. We do so by deducing a corresponding sea level pressure (SLP) from a surface pressure at each location. We add to the surface pressure the weight of a hypothetical air column underneath each location that extends to a reference sea level. A global SLP map can be thereby constructed.

Figure 2.3a shows that the winter (DJF) average SLP is as high as 1035 mb over central Asia and 1020 mb over central N. America, whereas that over the central part of Pacific and Atlantic is below 1005 mb. In contrast, Fig. 2.3b shows that the summer (JJA) average SLP

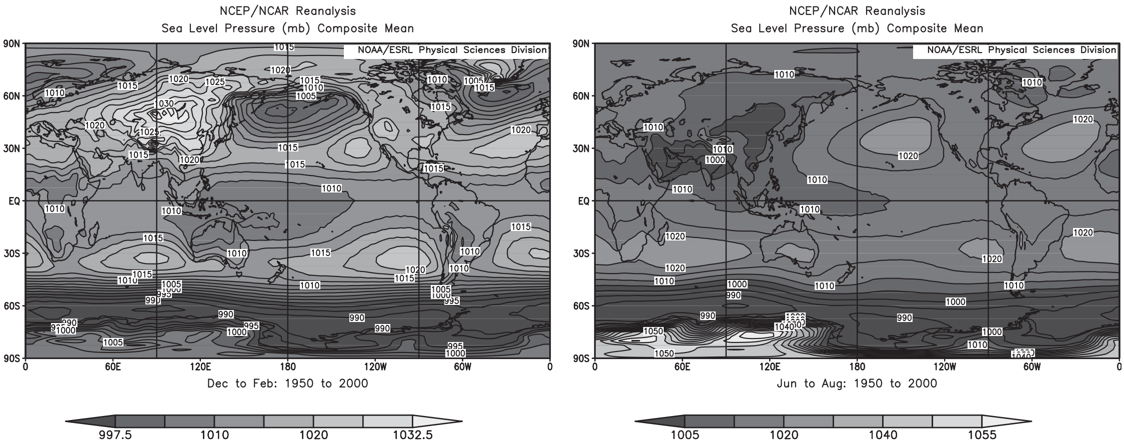

Fig. 2.3 Climatological mean sea level pressure over the globe in (a) winter (DJF) and (b) summer (JJA) in mb deduced from 50 years of NCEP/NCAR Reanalysis data.

is as low as 1003 mb over Asia and about 1012 mb over N. America, with the corresponding values over the N. Pacific and Atlantic oceans reaching 1025 mb in the subtropical latitudes. In other words, there is a pronounced differential in the seasonal average SLP over continent versus ocean. This is a rudimentary feature of the global monsoon phenomenon. A similar pressure differential over land versus ocean is observed in the southern hemisphere (SH), albeit in much smaller magnitude because the continental masses in the southern hemisphere are considerably smaller. For example, the SLP over the Indian Ocean and South Pacific in DJF (the summer season of SH) is noticeably lower than that over Africa, Australia and South America. In JJA, the area of maximum SLP moves over to the land masses. An elementary account for this broad feature can be given on the basis of hydrostatic balance.

Monsoon is a multifaceted complex phenomenon associated with the continent–ocean contrast ultimately due to the seasonal variation of the solar heating. The SLP distribution is naturally an integral feature of monsoon. Without going into the details of the monsoonal circulation, we can gain some appreciation of its rudimentary feature with the notion of hydrostatic balance and qualitative consideration of the net diabatic heating/cooling. Let us visualize what happens in the atmosphere over a continent and its adjacent oceans as a season begins to change. By inference, there is practically no SLP differential between a continent and ocean in a transition season, say the autumn (SON). As the winter season begins in a hemisphere, the solar radiation reaching the Earth's surface per unit area substantially diminishes day by day. The Earth nevertheless continues to emit longwave radiation to space nearly at the same pace. This would give rise to a net deficit in the surface radiation budget. The air just above the surface would therefore begin to drop in response. But since an ocean is an enormous heat reservoir, its surface temperature would not decrease nearly as much as the land surface temperature. Consequently, the air temperature over the land surface would drop faster and become greater than that over the ocean surface. Since an air parcel shrinks when its temperature drops, we would expect a greater reduction of the thickness of a shallow surface layer over land than over ocean. As this layer shrinks, less mass of air would remain in the column above a particular level. Recall that under hydrostatic balance, the pressure at a level is equal to the weight of all the air parcels above that level. Thus, the pressure at higher levels in winter would be lower over a continent than over an ocean. The corresponding horizontal pressure differential would induce a flow of air from ocean to land. The convergence of air towards an atmospheric column over land would in turn increase the SLP over land. There should be at the same time a divergent outflow from the land to the ocean near the surface because of a surface pressure differential in the opposite direction. However, under the influence of surface friction, the surface divergent outflow would be weaker than the convergent flow aloft. The net result would be a net build-up of mass of air in an atmospheric column and hence a net increase in the surface SLP. This process would continue throughout the first half of the winter season leading to a progressive build-up of the surface SLP over a continent. This is the fundamental reason why we observe distinctly higher mean SLP over a continent than over the adjacent oceans in winter. The larger a continent is, the greater would be such surface pressure differential as observed over Asia and N. America.

In summer, all processes mentioned above occur in a reverse sense. The qualitative account for the observed lower SLP over continents than over oceans in summer is the same. But, there is an additional complication in summer. The expected inflow towards a continent from its nearby ocean surface brings in a lot of water vapor. Such an influx of water vapor eventually precipitates somewhere over the continent. There is consequently an important diabatic heating as an integral part of the summer monsoon. The copious monsoonal precipitation greatly complicates the dynamics of the circulation. Nevertheless, the surface pressure differential between continent and ocean may be expected even without explicitly taking into account the moist dynamics.

2.2.2 Isobaric coordinates and governing equations

Since pressure monotonically decreases upward in the atmosphere as long as hydrostatic balance is applicable, there is one-to-one correspondence between the pressure at a point and the elevation of that point. Pressure therefore can be used as a vertical coordinate in an analysis. It is known as the *isobaric coordinate*. This coordinate is used in all operational weather forecast models. The advantage in using it is that meteorological instruments are designed to make measurements of weather data at a number of preset pressure values, e.g. 950, 900, 850, 700, 600, 500, …, etc. (mb). Such data can be directly used in a forecast model using an isobaric coordinate. This avoids the errors one would inevitably introduce if data at pressure levels are interpolated to height levels in a model using a height coordinate.

In trying to use pressure as a vertical coordinate in a model analysis, we need to transform all governing equations from height coordinates to pressure coordinates. Hydrostatic balance is invoked at the outset. Let us use local Cartesian coordinates as the horizontal spatial variables. We denote the height coordinates by $(\tilde{x}, \tilde{y}, \tilde{z}, \tilde{t})$ and the pressure coordinates by (x, y, p, t). The transformation relations are: $x = \tilde{x}$, $y = \tilde{y}$, $p = \tilde{p}(\tilde{x}, \tilde{y}, \tilde{z}, \tilde{t})$, $t = \tilde{t}$. The height of pressure surfaces, $z(x, y, p, t)$, is a dependent variable in an isobaric