from Einsele (1992)
p. 313-320

8 Subsidence

8.1 General Mechanisms Controlling Subsidence

Isostasy

Substantial sediment accumulation and the formation of sedimentary basins result from crustal subsidence. At least in the beginning of basin formation, tectonic subsidence must predate sedimentation, whereas later, subsidence may also be actively driven by an increasing sediment load (total subsidence). In this chapter, the different mechanisms leading to crustal subsidence as well as some models quantifying the development of subsidence versus time, i.e., subsidence history curves, are briefly introduced. A more comprehensive treatment of this topic is given by Allen and Allen (1990).

In most cases, tectonic subsidence of the land surface or sea floor is controlled by the principle of isostasy, thermal contraction of the lithosphere, and/or flexural loading (Fig. 8.1). According to the present view of *isostasy*, the elevation of the top of the crust is a function of the thicknesses and densities of several layers (sea water, sediments, solid crust consisting of igneous and metamorphic rocks, and solid upper mantle or mantle lithosphere, Fig. 8.1) resting on the viscous asthenosphere (mantle asthenosphere). Within the asthenosphere, *horizontal surfaces of constant pressure* can be assumed, which implies that the mass per unit area of the overlying rock column is everywhere the same. Hence one can write:

$$\rho_w h_w + \rho_s h_s + \rho_c h_c + \rho_m h_m + \rho_a h_a = \text{constant} \qquad (8.1)$$

where the layers (with thickness h) are sea water, w, sediment, s, crust, c, mantle lithosphere, m, and mantle asthenosphere, a. Density values, ρ, of these layers are listed in Fig. 8.1. The mass of the atmosphere per unit area is neglected in Eq. (8.1).

The base of the lithosphere is assumed to be a temperature boundary (approximately 1350 °C). Therefore, the hotter mantle asthenosphere has a somewhat lower density than the overlying mantle lithosphere. Thickening of the crust at the expense of the mantle lithosphere causes uplift of the land surface or sea bottom, whereas thickening of the mantle-lithosphere at the expense of the crust is followed by subsidence. For that reason, thick continental crust usually rises above sea level, whereas the top of much thinner oceanic crust lies several km below. Newly formed oceanic

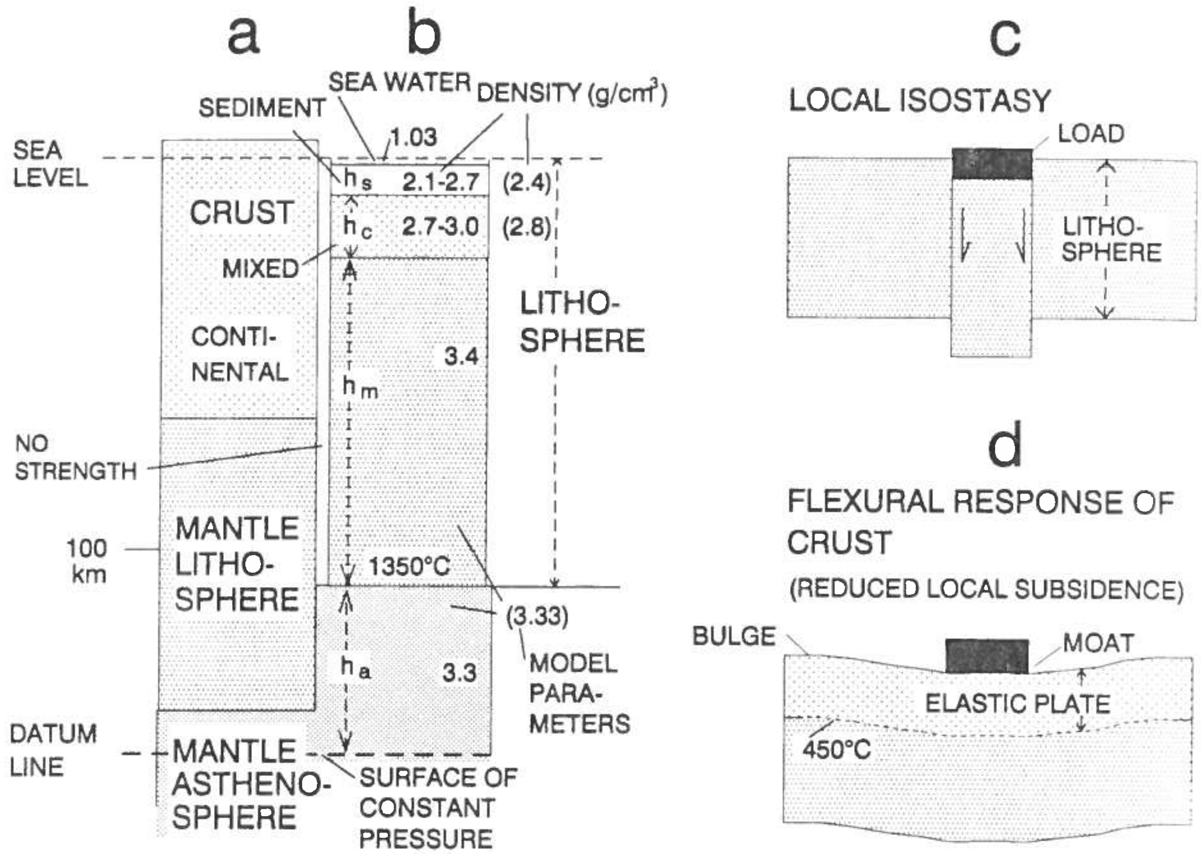

Fig. 8.1a-d. Principle of isostasy demonstrated by two crustal columns (**a** high plateau; **b** continental shelf) consisting of layers of different thickness and density on top of a surface of constant pressure. Values in *parenthesis* are used in the models (see text). **c** Local isostatic response of lithosphere to additional load. **d** Reduced but more widely extended subsidence due to flexural response of rigid crust

lithosphere is encountered at an average depth of about 2.5 km below sea level. This is also approximately the tectonic subsidence of a basin resulting from the partial replacement of thinning continental crust by upwelling denser upper-mantle material (Fig. 8.1b). Old oceanic crust on top of cooling mantle lithosphere may subside more than 6 km below sea level.

In order to quantify the mechanism of isostasy, two relationships have to be considered:

1. The thicknesses of the different layers on top of a given datum in the asthenosphere may change, but the mass of the total rock column per unit area will remain constant. Hence the sum of all changes in mass above the datum line is zero:

$$\Delta(\rho_w h_w) + \Delta(\rho_s h_s) + \Delta(\rho_c h_c) + \Delta(\rho_m h_m) + \Delta(\rho_a h_a) = 0 \qquad (8.2)$$

2. The changes in thickness of all layers above the datum line, including the thickness change, Δh_g, of the atmosphere, is zero.

$$\Delta h_g + \Delta h_w + \Delta h_s + \Delta h_c + \Delta h_m + \Delta h_a = 0$$

hence, the subsidence, S_i, of the land surface or sea floor (below the base of sediments) in relation to a fixed sea level or other datum line is

$$S_i = \Delta h_g + \Delta h_w + \Delta h_s \qquad (8.3)$$

The law of isostasy was first proposed by Airy; it is only valid for cases in which each crustal column behaves completely independently from its neighboring columns and does not support any adjacent loads. In addition, it as assumed that equilibrium conditions are established, which is not the case shortly after a rapid change in the load resting on the datum line within the asthenosphere. The type of isostacy under equilibrium conditions may be described as *local isostatic compensation*, in contrast to the flexural response of an elastic crust to a surface load (Fig. 8.1c and d).

Initial Subsidence of Water-Filled Basins (Without Thermal Effects)

In simplified cases, subsidence, S_i, due to local isostatic compensation, neglecting any thermal effects, can be calculated under the following assumptions: The densities of the different layers remain constant and the thicknesses of the layers (except that of sea water) are known or will be given as variables. Mantle lithosphere and mantle asthenosphere are combined into the thickness h_{ma} with the density ρ_{ma} (= 3.33 g/cm^3). Prior to subsidence, the surface of the crust is at sea level and covered neither by water nor sediment. For these conditions, one can simplify Eq. (8.2) to

$$\rho_w \Delta h_w + \rho_c \Delta h_c + \rho_{ma} \Delta h_{ma} = 0 \qquad (8.4)$$

and Eq. (8.3) to

$$\Delta h_w + \Delta h_c + \Delta h_{ma} = 0 \,,$$

$$\text{where } S_i = \Delta h_w \text{ (Fixed sea level)} \qquad (8.5)$$

After equating (8.4) and (8.5), one obtains

$$\Delta h_w = \Delta h_c \frac{\rho_c - \rho_{ma}}{\rho_{ma} - \rho_w} \qquad (8.6)$$

Inserting the density values listed above and in Fig. 8.1 (e.g., $\rho_w = 1.03$ g/cm^3 for sea water), we obtain

$$\Delta h_w = -\,0.23\,\Delta h_c \,, \text{ or } \Delta h_c = -\,4.34\,\Delta h_w \,.$$

If under these simplified conditions the crustal thickness h_c is reduced, for example, by 15 km (around half of its normal thickness), the top of the crust subsides from sea level to a depth of 3.5 km and creates space for a water body of the same depth. Similarly, the water depth of a basin can increase by 1 km, if the thickness of the mantle is increased by 4.34 km at the expense of the crust.

A more general expression for the initial subsidence of a basin filled with water up to the (fixed) sea level, derived from Eqs. (8.2) and (8.3), is (e.g., Suppe 1985):

$$S_i = \frac{(\rho_c - \rho_a)h_c + (\rho_m - \rho_a)h_m}{\rho_w - \rho_a}\left(1 - \frac{1}{\beta}\right), \qquad (8.7)$$

where ß = *stretching factor*, which describes the amount of stretching of a certain segment of the lithosphere. For z = original thickness and b = original width (Fig. 8.2), the new width after extension is increased to bß and the thickness reduced to z/ß. This formula, however, still does not take into account any sediment load or thermal effects which are usually associated with crustal thinning and the formation of a topographic low.

Inserting the values from the previous example (ß = 2, $\rho_m = \rho_a = 3.33$ g/cm^3, $h_c = 30$ km, and $h_m = 100$ km) into Eq. 8.7, the result is $S_i \approx 3.5$ km. With $\rho_m = 3.4$ g/cm^3, $\rho_a = 3.3$ g/cm^3 (Fig. 8.1), and the other parameters from this example remaining unchanged, the initial isostatic subsidence for the water-filled basin is $S_i \approx 5.5$ km.

These examples demonstrate that the results of such calculations strongly depend on the density values assumed for the mantle lithosphere and mantle asthenosphere, as well as on the thickness of the crust and lithosphere prior to stretching. Some authors have pointed out that stretching of a crust thinner than 18 km will generate uplift, whereas stretching of thicker crust causes subsidence. However, this boundary condition also varies considerably with the densities and thicknesses of the crust and mantle lithosphere (Jarvis 1984).

Initial Subsidence of Sediment-Filled Basins (Without Thermal Effects)

The *sediment load* of a partially or entirely filled basin causes additional subsidence due to isostatic compensation. This effect can be easily calculated if the former water-filled basin (water depth, h_w) is completely filled with sediments up to sea level (sediment thickness, h_s, and average density of sediments, for example, $\rho_s = 2.4$ g/cm^3). Then the elevation above the isostatic compensation depth remains the same as before, and from Eq. (8.3) we can assume:

$$h_s - h_w + \Delta h_a = 0 \,. \qquad (8.8)$$

Similarly, the mass above the compensation level does not change (Eq. 8.2), hence we obtain

$$\rho_s h_s - \rho_w h_w + \rho_a \Delta h_a = 0 \,. \qquad (8.9)$$

Equating (8.8) and (8.9) gives

$$h_s = \frac{(\rho_w - \rho_a)}{(\rho_s - \rho_a)}\, h_w \,, \qquad (8.10)$$

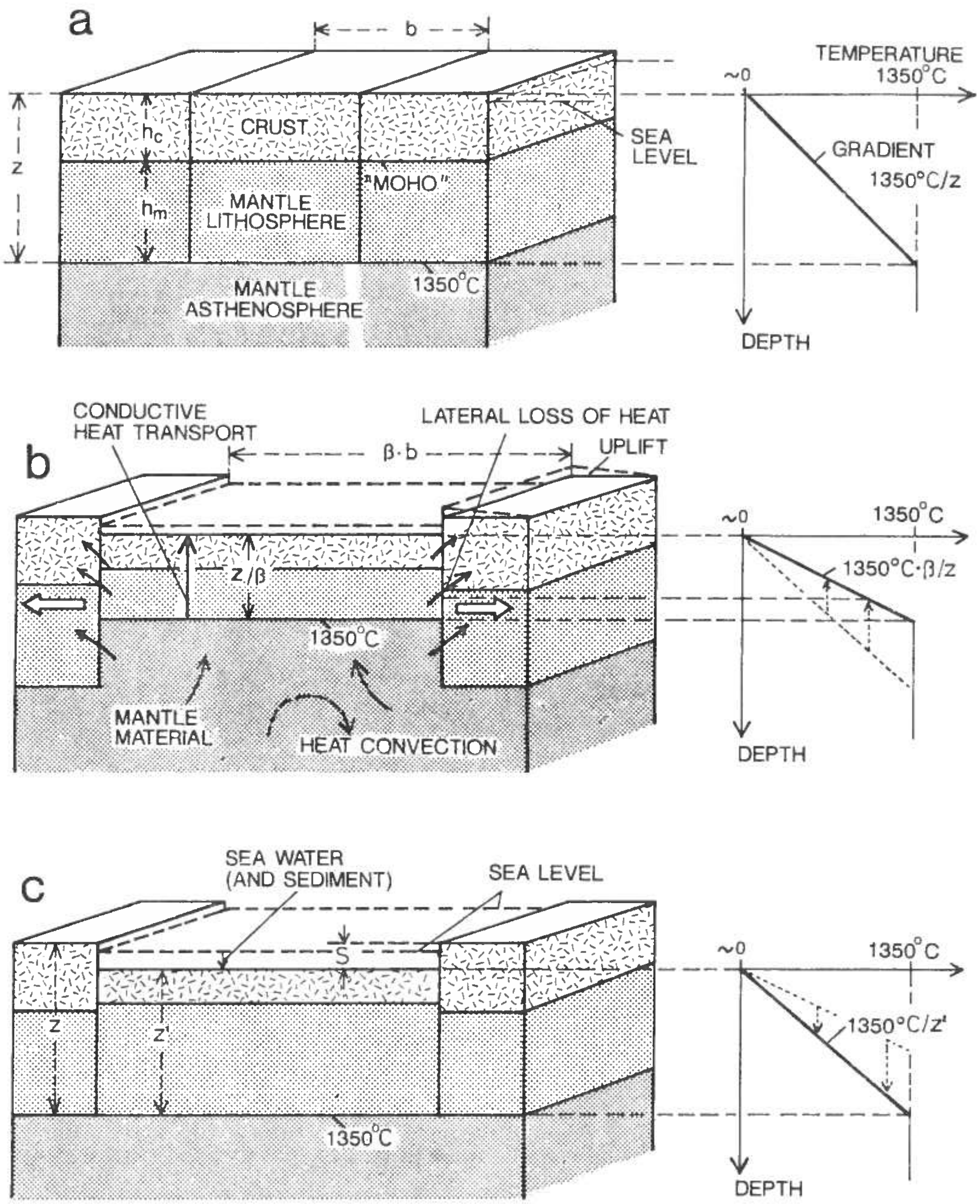

Fig. 8.2a-c. Finite-length extensional rift basin model. **a** Prior to rifting. **b** Initial subsidence due to isostatic adjustment at the end of short rifting event, buildup of high geothermal gradient. **c** Thermal subsidence due to slowly cooling lithosphere, geothermal gradient approximately reduced to original state. See text for explanation. (After McKenzie 1978)

where h_s is the maximum sediment thickness up to sea level.

With an average sediment density of $\rho_s = 2.15$ g/cm^3, a sediment thickness twice the initial water depth can develop (*amplification factor* 2); with $\rho_s = 2.55$ g/cm^3 the amplification factor is about 3.

In our example of a water-filled basin (based on Eq. 8.7) we obtained $h_w \approx 3.5$ km. Using the density values mentioned above and listed in Fig. 8.1, from Eq. (8.10) one gets $h_s \approx 8.7$ km. This signifies that a sediment-filled segment of this model basin will subside 2.5 times more than a purely water-filled segment, due to isostatic response of the crust. In comparison to an air-filled depression below sea level, the amplification factor for a corresponding sediment-filled basin becomes even greater (in our example approximately 1.46 x 2.5 ≈ 3.6, also see Jarvis 1984).

Thermal Uplift and Subsidence

If a column of the lithosphere is heated above the temperature of the surrounding rocks, its mass remains unchanged, but its volume increases and may cause uplift at the land surface or sea bottom. Upwelling hot mantle material, for example at oceanic spreading centers, generates oceanic ridges which later, due to cooling of the oceanic crust and thickening of the underlying mantle lithosphere, subside considerably with age by several km. Similar ly, upwelling hot mantle material below a thinning continental crust will first tend to raise the surface of the thinned crust, if this thermal expansion is not exceeded by the effects of isostasy, and then cause subsidence due to cooling. This type of subsidence is referred to as thermal subsidence, S_t. Thermal subsidence and uplift, apart from changes in temperature and geometrical factors, are controlled by the coefficient of thermal expansion (or contraction), $\alpha = 3.4 \times 10^{-5}/°C$, and the thermal diffusivity of the lithosphere, $k = 8 \times 10^{-7}$ m^2/s.

If for example the temperature of a 50 km high rock column is raised or lowered uniformly by 300 °C, its height will increase or decrease due to thermal expansion or contraction by ca. 0.5 km. This is, of course, an unrealistic example, which only demonstrates the order of magnitude of such thermal effects. In nature we observe a thermal gradient from the base of the mantle lithosphere (1350 °C, Fig. 8.2) to the land surface or sea water (slightly above 0 °C). With changing lithosphere thickness, this gradient and thus the temperatures at different depths vary, but the combined effect of this process causes approximately the same amount of expansion or contraction as that calculated for a rock column which is subjected to the same change in mean temperature as the lithosphere.

As a result of mantle upwelling, a steep temperature gradient is established. Later, this gradient is slowly reduced by heat loss to the atmosphere, until the original gradient will be restored (Fig. 8.2b and c). This process is described by the thermal diffusivity of the lithosphere. Cooling of an abnormally heated lithosphere takes tens of million years, and therefore thermal subsidence is a long-term, slowly decaying process. The mathematical treatment of thermal expansion and contraction of the lithosphere is beyond the purpose of this text (see, e.g., Royden et al. 1980; Watts and Thorne 1984; Sawyer 1988), but the following basin modeling does include thermal effects.

Flexural Response of the Crust to Loading

It is a well-known fact from geological observation that the lithosphere adjacent to local or regional loads (water, ice, sediment accumulation, isolated seamounts on top of oceanic crust) or unloading (lake dessication, ice melting, rock denudation) reacts by downwarping or uplift, respectively (Fig. 8.1 c and d). The best studied examples are continental ice caps, which cause not only isostatic compensation below the ice, but also some subsidence in areas adjacent to the ice load. Later, after ice melting, these regions begin to rise again in conjunction with the formerly ice-loaded crust. This phenomenon may affect a zone 50 to 300 km wide adjacent to the loaded and unloaded crust.

Several authors have made an attempt to simulate the flexural response of the crust to a given load (e.g., Watts et al. 1982; Watts and Thorne 1984). They treat the lithosphere either as an elastic or visco-elastic plate underlain by the viscous mantle. In the case of *pure elastic behavior*, the mechanics of lithospheric flexure can be compared with the bending of sheets or beams having a certain flexural rigidity. This flexural rigidity, D, in turn is a function of the elastic constants of the rock material, i.e., Young's modulus, E, and Poisson's ratio, σ, as well as thickness, h_e, of the elastic lithosphere. Then we obtain

$$D = \frac{E\, h_e^3}{12\,(1 - \sigma^2)} \quad . \tag{8.11}$$

Equation (8.11) clearly shows that the elastic thickness, h_e, is the most important factor controlling the rigidity or flexural strength of the lithosphere.

As has been derived from observations in nature, the rigidity, D, of the lithosphere may vary by four orders of magnitude (Fig. 8.3). It is primarily a function of the plate age at the time of loading. Young oceanic crust at or near a mid-oceanic ridge has the lowest rigidity, while old continental crust, for example, below the fill of foreland basins or Precambrian shields under an ice load, show very high values. The base of the elastic plates is defined by the depth of the 450 °C isotherm (cf. Fig. 8.15). Thus, plate thickness, h_e, is relatively thin during or shortly after a major thermal event, but increases substantially with age after this event. The mechanical properties of both oceanic and continental lithospheres appear to be similar (Karner et al. 1983).

Furthermore, the vertical displacement of the bending crust from its original elevation is affected by the densities of the underlying viscous mantle material and, if present, the overlying water body or sediments (see, e.g., Bott 1982; Watts et al. 1982). The resultant of these two effects is an upward pressure or an additional buoyant resistance of the asthenosphere to the bending.

The mathematical treatment of these combined mechanisms for locations at various horizontal distances away from the load is rather complicate. In addition, the load on the plate can be distributed nonuniformly. Therefore, only some results of such modeling can be discussed here.

In the case of a relatively small load of limited areal extent (Fig. 8.4a), an elastic lithosphere can support this surface load by distributing it onto a larger area. Hence, subsidence is less than that of a locally compensated load, and bending of the lithosphere is not very pronounced. For greater loads with an extent wider than the thickness of the lithosphere (Fig. 8.4b) and long loading times, at least the center of the load tends to approximate isostatic equilibrium and the flexural response of the adjacent crust can become very important. The time necessary to reach such an equilibrium depends on the rheological properties of the lithosphere (see below). The initial response of the lithosphere to an instantaneously applied load is a rapidly downwarped flexural moat around the load. In general, the vertical deflection decreases in a sinusoidal manner away from the load. The mass of material displaced from beneath the downwarped basin must approximately equal the mass of the applied load. Beyond the zone of decaying subsidence, an upwarped peripheral bulge is usually developed (Fig. 8.4). Where old, thick, elastic lithosphere is being flexed, the width of the flexure is much broader than that of a young or recently heated thin plate, provided the load and its geometry are about the same.

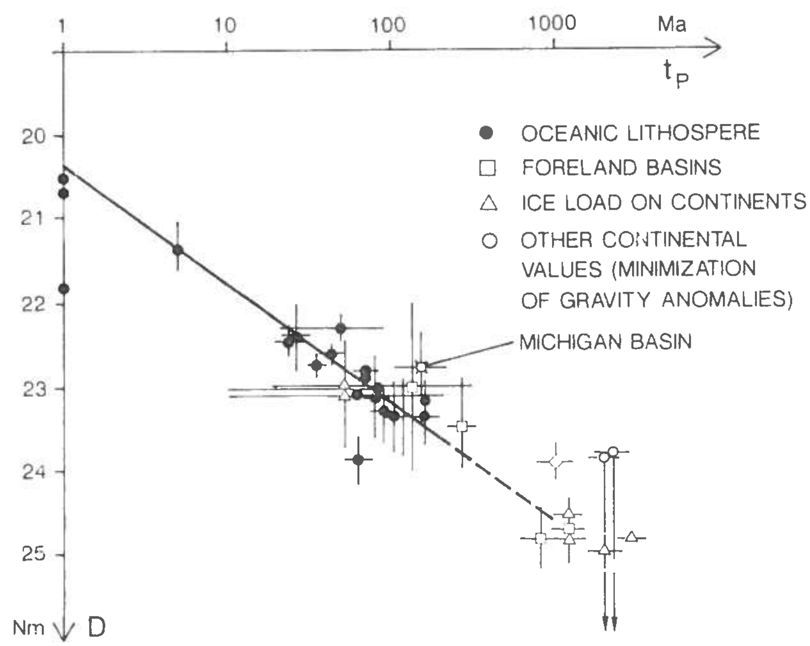

Fig. 8.3. Effective elastic rigidity, D, of oceanic and continental lithosphere in relation to plate age, t_P, at the time of loading after a major thermal event. (After Karner et al. 1983)

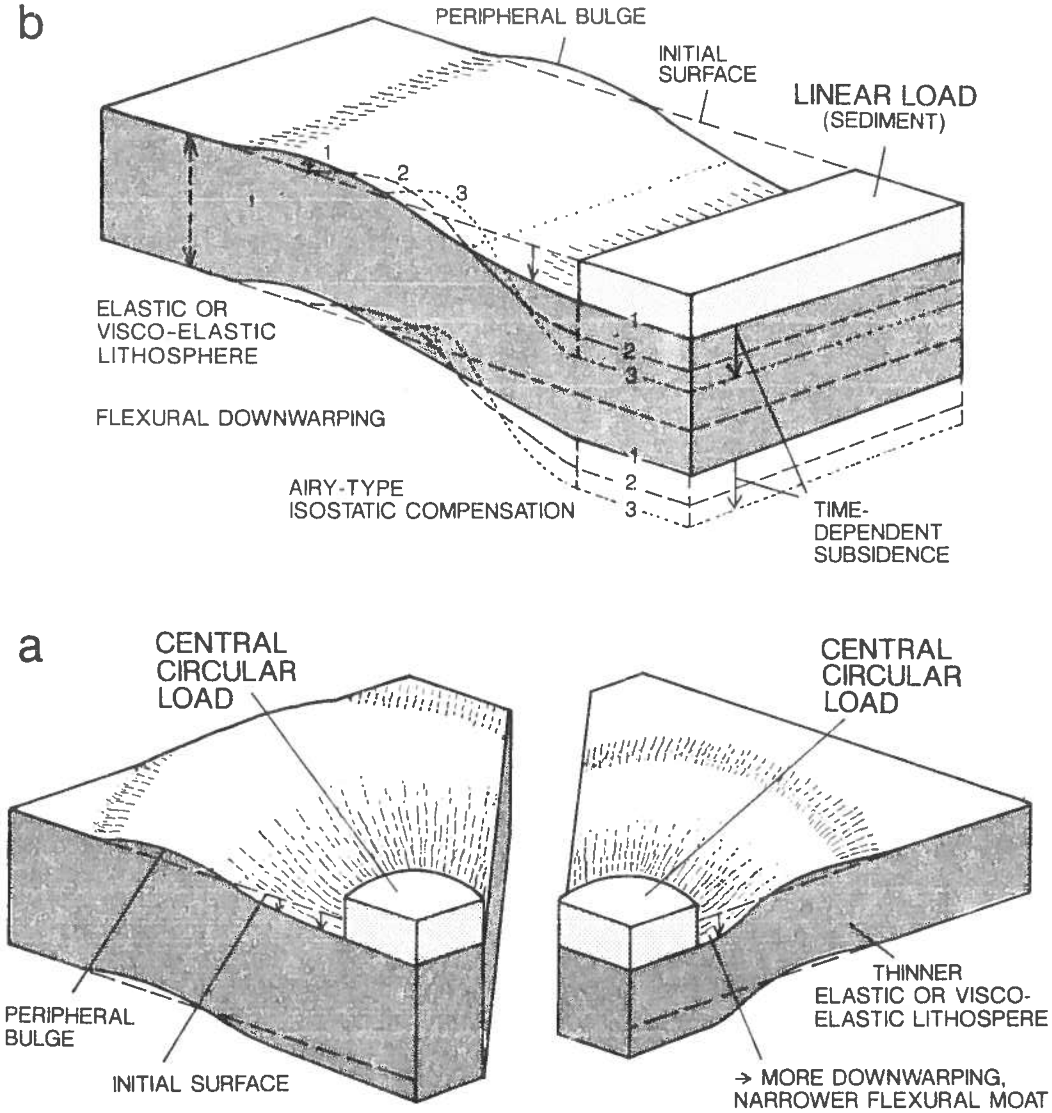

Fig. 8.4a,b. Flexural response of lithosphere adjacent to (**a**) local small or (**b**) large, wide linear load, not to scale. **a** A load on top of relatively young, thin lithosphere creates a deeper, narrower flexural moat than a load on thick, old lithosphere. **b** Under a long-persisting load, rapid initial subsidence *(1)* may be followed by further slow subsidence *(2)* until the load is ultimately compensated by local isostasy *(3)*. Simultaneously, the peripheral bulge migrates toward the load. (Partially based on Quinlan and Beaumont 1984).

Aside from such general rules, the results of modelling the lithospheric flexure often remain uncertain and even problematic. The flexural rigidity of the lithosphere appears to be dependent not only on plate age at the time of loading, as mentioned above, but also on the duration of loading (e.g., Bott 1982). An initially high rigidity may decrease with time following loading. Short-period loading (on the order of 10 000 years) generates less bending of the lithosphere than loading times of the order of 100 Ma. This phenomenon may be related to thinning of the elastic plate with age due to a rising 450 °C isotherm under the long-persisting load, or it may be ascribed to a visco-elastic lithosphere and a softening of the crust with time.

If the load remains in place for a long time period, deformation of the lithosphere changes in a manner as shown in Fig. 8.4b. After a first stage of rapid downwarping, lithospheric material at depth under the load relaxes stress, thereby creating a deeper central depression in conjunction with a narrower basin. The peripheral bulge is progressively uplifted and migrates toward the load. Ultimately, relaxation will evolve toward a state of local isostatic equilibrium.

Such a time-dependent development is primarily controlled by the mechanical state of the lithosphere. Although some workers strongly believe that models based on elastic condition plates can fairly well predict the processes observed in nature, others favor a uniform visco-elastic (Maxwell) model of the lithosphere. A third possibility is the assumption of a Maxwell layer with temperature-dependent viscosity overlying a low-viscosity fluid (e.g., Quinlan and Beaumont 1984). Few of the models have taken into account additional local heat sources associated with special magmatic or volcanic events during the history of a continental margin or other flexure-influenced basins.

Summary of Factors Controlling Subsidence

In summary, subsidence of the floor of a sedimentary basin (or uplift of the land surface) is controlled by the following principal factors:

- Thinning (or thickening) of the lithosphere due to horizontal extension (or compression, underplating).
- Upwelling of mantle material in response to crustal thinning.
- Increase (or decrease) in lithospheric density, for example due to
  1. cooling (or heating),
  2. pervasive dike intrusion or other injections of magma,
  3. phase changes in the crust such as melting and magma crystallization as well as transitions of minerals of relatively low density into minerals of higher density (or vice versa).
- Isostatic subsidence (or uplift) in response to sediment loading (or erosion).
- Flexural loading.
- Subcrustal magma convection.

The subsidence models of Chapter 8.3 have been worked out for relatively simple cases and deal primarily with the effects of isostasy, including sediment loading and thermal subsidence as well as flexural downwarping. The results obtained for simplified model basins can be compared with subsidence curves observed in nature for characteristic basin types of similar tectonic setting.

8.2 Methods to Determine Subsidence of Sedimentary Basins

Introduction

The *subsidence history* of sedimentary basins may be described in different ways. Structural geologists and geophysicists are primarily interested in that part of subsidence which is controlled by crustal processes (tectonic or thermo-tectonic subsidence), whereas sedimentologists and stratigraphers deal with the progressive burial history of sediments (total or cumulative subsidence) and often want to interpret subsidence curves in terms of sedimentation rates, paleo-water depth, and sea level changes.
Proceeding from endogenetic, geodynamic to exogenetic processes, the subsidence history at a certain location within a basin is controlled by the following mechanisms (Fig. 8.5):

- *Thermo-tectonic subsidence* due to processes within the crust and the mantle lithosphere (extension, thermally induced changes in thickness and density, see below). The resulting, essentially exponential subsidence curve (A) for an air-filled basin represents the minimum amount of subsidence which is not affected by any other mechanism (such as filling by water or sediment, see below).
- *Subsidence caused by water load*. When the tectonically driven, subsiding basin is filled up to sea level with water, the amount of subsidence versus time increases considerably (curve B in Fig. 8.5).
- *Subsidence caused by sediment load*. If a tectonically subsiding basin is permanently filled with sediments up to a fixed sea level, subsidence in relation to a water-filled basin is magnified by a factor about 2.5 (see above), but the resulting subsidence curve (C) is still exponential and smooth.

The regular subsidence curve for a sediment-filled basin can be modified by several minor factors:

1. At certain times, the *basin is not completely filled with sediments*. Hence, the total load on top of the basin floor is reduced according to the paleo-water depth, and subsidence decreases (Fig. 8.5, Curve D). Occasional erosion of sediment leads to additional unloading with similar effects (Curve E).