

Sedimentary Basins

Revision Material

This handout lists the topics covered in the two lectures on sedimentary basins and provides a few key diagrams. Either of the following two books may be useful for revision and both can be found in the College Library.

- *The Solid Earth: An introduction to global geophysics*, CMR Fowler¹.
- *Basin Analysis: Principles and Applications*, P Allen & J Allen.

Also, there is a vast amount of revision material available on the internet; try typing some of the keywords below into a search engine.

Introduction

What are sedimentary basins? Sedimentary basins are regions where considerable thicknesses of sediments have accumulated (in places up to 20 km). Sedimentary basins are widespread both onshore and offshore. The way in which they form was a matter of considerable debate until the last 20 years. The advance in our understanding during this very short period is mainly due to the efforts of the oil industry.

Probing sedimentary basins The cheapest and most effective method for probing large regions of the subsurface is seismic reflection profiling (recall lecture on 2D, 3D and 4D seismic reflection profiling).

¹Section numbers given in this document refer to the 2nd Edition, published in 2005

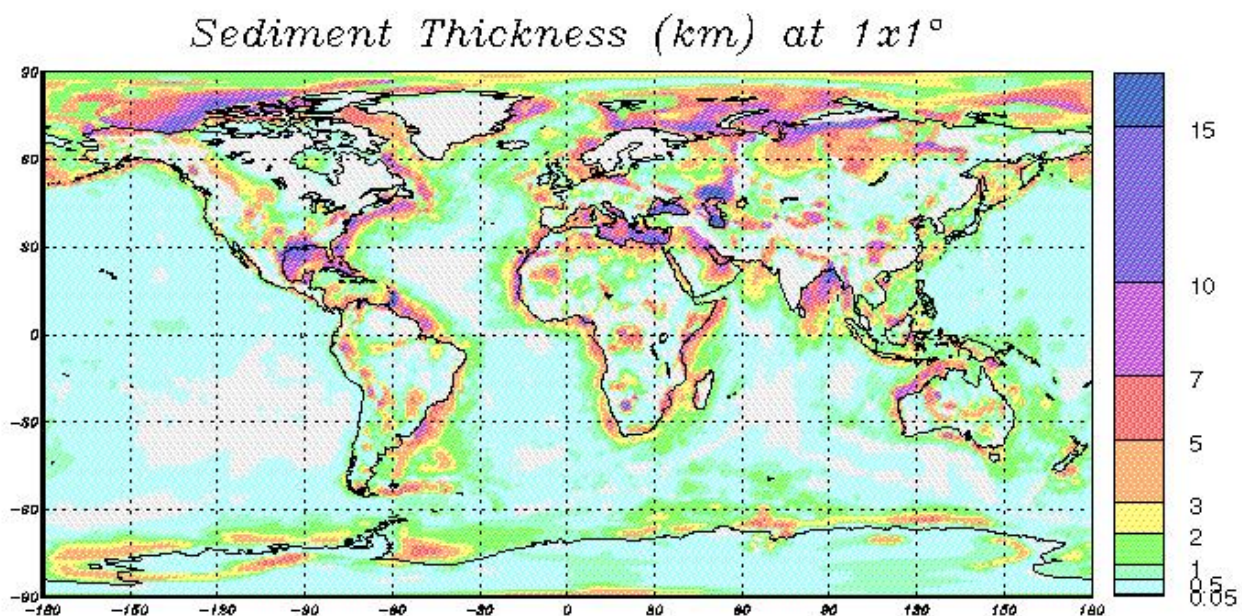


Figure 1: Global sediment thickness map

Where basins cover onshore regions, surface geological mapping is useful, especially when combined with seismic profiling to build a detailed 3D picture. Drilling is more expensive but necessary to prove, evaluate and produce discoveries.

Basin classification schemes Extensional basins, strike-slip basins, flexural basins, basins associated with subduction zones, mystery basins. There are many different classification schemes for sedimentary basins but most are unwieldy and use rather spurious criteria (see for example Basin Analysis §1.4). The most useful scheme (presented here) is very simple and is based on basin forming mechanisms. About 80% of the sedimentary basins on Earth have formed by *extension* of the plates (often termed *lithospheric extension*). Most of the remaining 20% of basins were formed by *flexure* of the plates beneath various forms of loading (this class will be covered in the next lecture). *Pull-apart* or strike-slip basins are relatively small and form in association with bends in strike-slip faults, such as the San Andreas Fault or the North Anatolian Fault (see lecture on Earthquake Prediction). Only a very small number of basins still defy explanation, although we suspect that at least some of these have a thermal origin (Solid Earth §10.3.4).

Extensional Basins

Relationship between extensional basins and ocean basins Basin stretching models follow logically from the thermal model used to account for the subsidence of oceanic plates on either side of mid-ocean ridges. Recall that the oceanic plate thickness is very small at mid-ocean ridges, where diverging plates draw up hot mantle close to the surface. But plates always want to be at their equilibrium thickness of around 120 km because this represents a state of thermal equilibrium. As the plates thicken, their average density increases and the principle of isostasy (i.e. Archimedes' principle) means that they subside to an equilibrium depth, the subsidence being exponential in form.

Deformation styles Extension of a plate is accommodated by brittle deformation in the upper part of the plate (i.e. normal faulting; recall faulting lecture) and plastic deformation in the lower part of the plate. The plate stretching model itself arose from observations of crustal thinning, subsidence, heatflow and normal faulting in the Aegean, a region of active extension. Evidence for normal faulting in the Aegean is as follows.

1. *Fault breaks* and *fault planes* are observed in the field.
2. *Uplift*, *subsidence* and *rotation* related to normal faulting can be inferred from geological indicators of past sea-level and geomorphological indicators of past horizontal land surfaces.
3. Earthquake analysis from global seismology. Fault plane solutions (plotted as beach ball symbols on maps) allow remote determination of the sense of movement on the fault (i.e. normal, reverse, strike-slip). See The Solid Earth §4.2.8.
4. Geodetic measurements. GPS surveying allows both extension and subsidence/uplift across faults to be mapped to sub-cm scale (e.g. The Solid Earth Figure 4.21).

The lithospheric Stretching Model The lithospheric stretching model advanced by McKenzie in 1978 is now one of the most widely cited papers in Earth Sciences. The model has three stages, illustrated in Figure 1 and discussed in The Solid Earth §10.3.6 and Basin Analysis Chapter 3.

1. The *pre-rift* phase. Undeformed lithospheric template.
2. The stretching (or rifting or *syn-rift*) phase. Rapid extension of the continental plate causes it to thin, and hot mantle wells up passively beneath. Block faulting and rapid subsidence are associated

with this phase. The key parameter is β , the length to which a unit length of continental plate is stretched.

3. The cooling (or *post-rift*) phase. Once stretching has ceased, the continental plate cools and re-thickens to achieve thermal equilibrium (analogous to the oceans). This phase is associated with *further subsidence*, exponential in form, but there is *no faulting*.

Testing the lithospheric stretching model

1. Measure the thickness of the crust across the basin using wide-angle seismic profiling. The crust should be thickest beneath the basin margins and thinnest beneath the basin centre.
2. The total extension accommodated by a normal fault is termed the *heave*. It is measured as the horizontal separation between two points that were originally in contact before the fault cut the horizon. These measurements can be made on interpreted seismic reflection profiles. Summing the heaves for all normal faults along a transect across a basin provides an estimate of total extension across the basin.
3. Sediment geometries. Pre-rift sediments should not change in thickness across a fault. Syn-rift sediments should thicken into the hanging wall (subsiding side) of a normal fault. On the footwall (upthrown side) syn-rift sediments should thin or erosion should occur. These observations can be made from field mapping or interpretation of seismic reflection profiles. Post-rift sediments are unfaulted.
4. Regional subsidence occurs across the basin as an isostatic response to changes in crustal and lithospheric thickness. The theoretical subsidence history can be determined by calculating the thermal evolution of the lithosphere, given the total extensional strain β and the duration of the rifting period (e.g. Basin Analysis Figure 3.12). The real subsidence history of a basin can be reconstructed from borehole records or interpreted seismic profiles. Sediment accumulation can be tracked using the technique in Practical 4 of this course, to find the subsidence history of the floor of the basin. Subsidence histories reflect both lithospheric stretching, which forms the basin, and the mass of sediment deposited, which modifies the basin shape. The effect of sediment mass

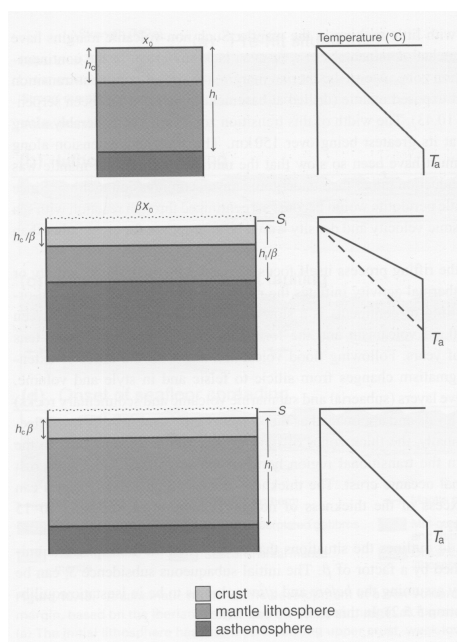


Figure 2: Instantaneous lithospheric stretching model, from The Solid Earth Figure 10.44, after McKenzie (1978). At time $t = 0$ the continental lithosphere has initial length x_0 and thickness h_1 (top). It is instantaneously stretched by a factor β (middle). By isostatic compensation, hot asthenospheric material rises to replace the thinned lithosphere. The temperature of the lithosphere is assumed to be unaffected by this stretching, but the stretched lithosphere is out of thermal equilibrium (dashed line). The stretched lithosphere slowly cools and thickens until it finally re-attains its original thickness (bottom). An initial subsidence S_i occurs as a result of the replacement of light crust by denser mantle material. Further subsidence occurs as the lithosphere slowly cools. The final subsidence is S .

is removed using a standard technique known as *backstripping* to yield a subsidence curve for an equivalent basin filled entirely with water. Observed and theoretical subsidence curves can then be compared.

- Heatflow within a basin is predicted by the lithospheric stretching model (e.g. Basin Analysis figures 3.12 & 3.13). Heatflow increases as during the syn-rift phase because the lithosphere is thinned and decreases exponentially during the post-rift phase as the lithosphere re-thickens. Theoretical heat-flow curves can be compared with heatflow measurements determined from borehole information.

Passive continental margins When stretching proceeds to $\beta = \infty$, the continental plate breaks apart and a new mid-ocean ridge plate boundary is born. On either side of the new ocean basin, the new oceanic plate is attached rigidly to the continental plate. These boundaries are known as passive continental margins. Deep water passive margins are important sites for hydrocarbon exploration, now that hydrocarbon reserves in shallow water basins are being used up.

Flexure of the Lithosphere

Strength of the Lithosphere Although about 70% of the sedimentary basins that have formed on this planet during the past 600 million years have done so by lithospheric extension, there are still several other important types of basin. Foremost amongst these are basins which form as a result of plate flexure or bending. We have seen that Airy isostasy (i.e. the geological application of Archimedes' Principle) is often a very useful approximation. Airy isostasy assumes that the plate is weak, so that any 'load' placed upon it (i.e. any change in thickness of the crust and/or lithospheric mantle) is supported by

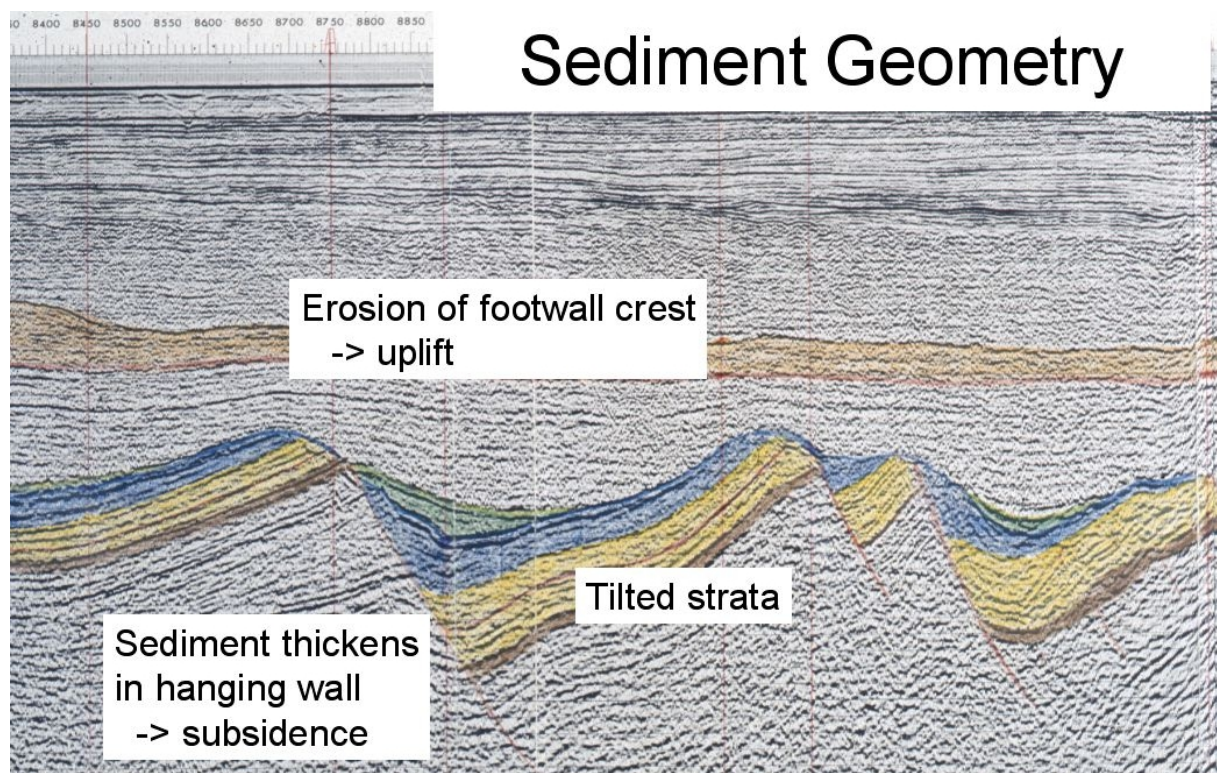


Figure 3: Sediment geometries associated with normal faulting, from an interpreted seismic reflection profile across the northern North Sea extensional basin.

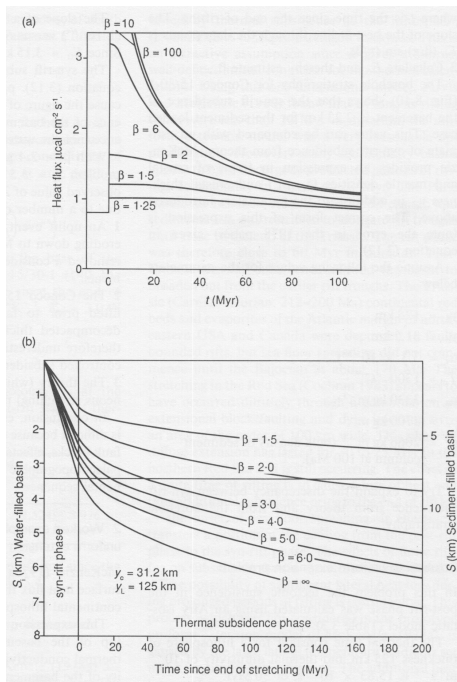


Figure 4: Subsidence and heat flux predicted by the lithospheric stretching model, from Basin Analysis Figure 3.12, after Dewey (1982). (a) Heat flux as a function of time for various values of the stretch factor β . (b) Subsidence corresponding to the heat flux patterns in (a). The rate of thermal subsidence decreases with time. The curves shown refer to a lithosphere with an initial thickness of 125 km and an initial crustal thickness of 31.2 km.

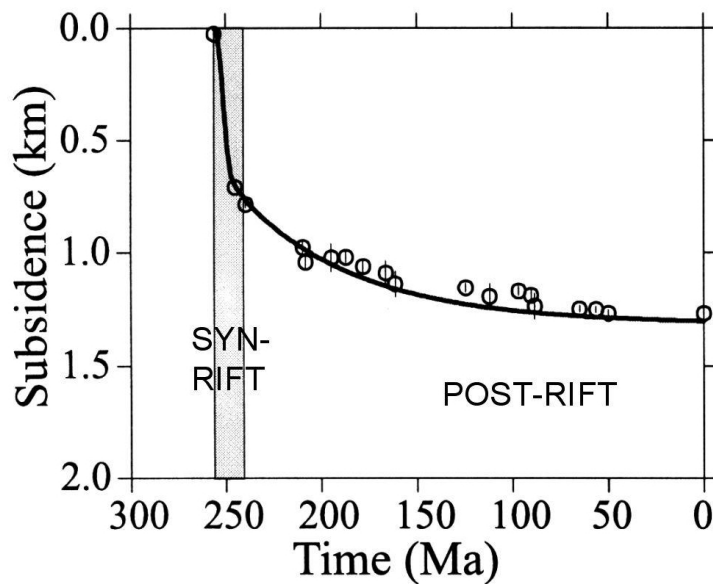


Figure 5: Observed subsidence of the southern North Sea extensional basin (dots). Theoretical subsidence curve calculated using the lithospheric stretching model that best fits the observations (line).

vertical deflections in the region where the load is applied and not in adjacent areas. However, there are important situations where this principle fails to operate and the strength of the plate plays a significant role. Clearly if the plate is very strong and the load is relatively small, there will be no deflection of the plate. In between, we envisage a situation of load support where small loads are supported mainly by the strength of the plate but larger and larger loads are supported by greater and greater degrees of isostatic compensation. In these cases, any load placed on the plate will cause bending of adjacent areas: these are the flexural basins where sediments can accumulate. The widths of such basins are governed by the flexural strength of the plate and their depths depend mainly upon the nature of the load. See The Solid Earth §5.7.

Flexural loading in the oceans Geophysicists recognise two types of flexural loading in the oceans. The first type is exemplified by Hawaii, which is a volcano around 9 km high whose tip pokes out only a few km above sea-level. This enormous volcano represents a load on the plate. A flexural moat surrounds the volcano and the moat is flanked by a peripheral bulge. The Hawaiian situation has been modelled by loading of an unbroken elastic plate by a line load. The other type of flexural loading is associated with subduction zones. In this case, the load is the old, cold, dense oceanic plate sinking back down into the mantle. Once again, a flexural depression lies next to the subduction zone and this depression is flanked by a bulge. Flexure at subduction zones is modelled by end-loading of a broken elastic plate. (The Solid Earth figures 5.14–5.16).

Flexural loading of the continents The most important examples of flexure of continental plates are associated with continent-continent collision. When such a collision occurs, regional shortening leads to thickening of the crust, and this thickening represents the load. Crustal shortening and thickening is accommodated by thrust faults in the zone of brittle deformation at the top of the plate. The flexural depressions either side of the load are known as *foreland basins*. Good examples of this class of basin have developed on either side of the Alps (the Molasse Basin and the Po Basin) and either side of the Himalayas (the Tarim and Ganges Basins). See The Solid Earth §10.3.5 and Basin Analysis chapter 4.

The Fate of a Sedimentary Basin

Isostasy The key to what happens to a basin subsequent to its formation is isostasy. Foreland basins (and flexural basins in general) forfeit their *raison d'être* once the loading forces which depress the plate have been removed. In other words, if thrust faulting ceases and the thrust sheets are eroded away, the foreland basin will be rapidly uplifted as the plate regains isostatic equilibrium. Note that erosion is an important form of unloading which itself drives further uplift. Clearly, foreland basins are rather fragile affairs; although many can be seen at present, there are few good examples in the geological record. Basins formed by extension of the plates are quite different since isostatic equilibrium is maintained throughout their development, from the rifting stage right through to the end of thermal subsidence. As a result, basins like the North Sea will simply sit there until something ghastly happens to them (e.g. they are squashed).

Orogenic uplift The uplift and erosion of sedimentary basins is called *basin inversion*. The term is most commonly applied to extensional sedimentary basins which have been shortened, showing normal faults which have been reactivated as thrust faults. The southern North Sea, southern England and the Celtic Sea offshore south of Ireland are good examples of inverted sedimentary basins. Much of the Alpine mountain chain consists of compressed extensional passive margins which were originally associated with the development of the Tethys Ocean several hundred million years ago.

Epeirogenic uplift Besides shortening caused by horizontal plate movements (known as *orogeny*), there is another important way in which sedimentary basins become uplifted and eroded (known as

epeirogeny). Although the mantle beneath the plates is solid, it convects as a fluid over geological timescales (recall mantle lecture). When a rising mantle plume hits the base of a plate, it deflects the plate upwards over a region more than a thousand kilometres in diameter. Heat within mantle plumes also generates melting, leading to centres of volcanic activity known as *hotspots*. The best examples of hotspots overlying mantle plumes are Hawaii and Iceland.