6. BASIN CLASSIFICATION AND SUBSIDENCE MECHANISMS

Having discussed the primary mechanisms of subsidence we can briefly focus on the plate tectonic settings of major sedimentary basins and examine their typical subsidence histories and mechanisms (Fig. 6.1). Much has been written about the driving mechanisms of basin formation in most tectonic settings. An early overview was provided by Dickinson (1976). The purpose of this chapter is not to attempt to summarize the state of knowledge of basin evolution. Instead, we simply define each basin type, following the basin classification scheme of Dickinson (1976), and focus on a few key points regarding the mechanisms of basin evolution. Due to space limitations we do not cite all of the relevant literature but provide just a few key references.

The subsidence curves (Fig. 6.1) come primarily from the published literature, augmented by analyses done by ourselves or by students in our sedimentary basins course at the University of Wyoming. All the curves have been backstripped following the local-isostatic method described above (Steckler and Watts, 1978). However, inconsistencies arise from the use of different time scales, compaction corrections and paleowater depth estimates made by the various authors. Nonetheless, the overall consistency of the subsidence curves in each of the various tectonic settings suggests that use of different scales by different workers do not generate errors large enough to mask the overall trends.

A. PASSIVE MARGINS AND RIFT BASINS

Atlantic-type margins and back-arc basins basically form by crustal extension and formation of an oceanic basin. Following Salveson (1978), the evolution of passive margins can be broken into five stages (Fig. 6.2) that explain many aspects of the preserved sedimentary record. His model assumes symmetry of the rift about its axis as is seen in some Atlantic margins (Keen et al., 1989); however, asymmetric rifting by development of a half graben may also occur (Lister et al., 1986). The subsidence of Atlantic-type margins have been well studied (see overview in Watts, 1981) and consists of an early extensional stage during which initial subsidence of rift basins takes place as an isostatic response to crustal thinning; followed by a post-rifting phase of subsidence that is driven by thermal reequilibration as the mantle lid cools and the lithosphere thickens. McKenzie (1978) has proposed the most popular model for basin subsidence in these settings, which emphasizes that as long as extension is instantaneous (i.e., less than 20 My) the initial subsidence will be complete before subsidence due to conductive heat loss begins. Uplift of the shoulders of the rift (Fig. 6.2) may be the result of two-layer extension, where extension of the mantle lid is distributed over a broader area than the crust (Rowley and Sahagian, 1986; Royden and Keen, 1980); lateral heat loss along the margins of the rift (Watts, 1981; Watts et al., 1982); and isostatic adjustment due to the geometry of bounding normal faults (Braun and Beaumont, 1989; Taber, 1927).

We have discussed examples of extensional/thermal subsidence throughout the course. Some recent to ancient passive margin settings are shown in Figure 6.1. In the case of the eastern U.S. Atlantic margin and the early Paleozoic Cordilleran miogeocline of western North America, the early stages of subsidence are not well constrained (shown with dashed lines), either because of poor age dating and/or poor exposure. However, Tertiary rifting of the Gulf of Lion (Steckler and Watts, 1980) indicates rapid initial subsidence as suggested by McKenzie (1978). In any case, the thermal parts of the subsidence curve mimic that of the subsiding ocean floor (shown for comparison on the same graph).

B. TRANSFORM BASINS

Pull-apart basins (Fig. 6.3) associated with strike-slip faults share some characteristics in common with passive margin sequences. The major differences are that pull-apart basins commonly do not go to completion (i.e., to the point where oceanic crust forms) and that pull-apart basins are much smaller and shorter lived features (Aydin and Nur, 1982; Karner and Dewey, 1986; Mann et al., 1983). This latter point is important in terms of their subsidence history. As pointed out by Pitman and Andrews (1985), the small dimensions of these basins allow heat to be removed, in part, by lateral conduction from the upwelled asthenosphere into the relatively cool adjacent mantle (Fig. 6.4). The result is that, in contrast to passive margin sequences that lose heat primarily by vertical conduction, pull-apart basins undergo thermal heat loss and subsidence at the same time as the basin is isostatically subsiding due to crustal thinning during extension. The resulting subsidence curves from transform basins (Fig. 6.1) contain very rapid, nearly linear, subsidence during fault movement and basin extension. The subsidence curves tend to have short tails of somewhat slower subsidence, probably representing cooling of the small remaining thermal anomaly once the fault ceases to be active (Fig. 6.4).

The fact that many of these basins are short lived, probably reflects the short duration that consistent motion exists over any small segment of a transform fault (Mann et al., 1983). After a few million years either the fault configuration changes so that a transtensional segment becomes transpressional or fault movement ceases and relative motion begins along a new fault trace.

C. FORELAND BASINS

These much studied basins form by flexure of the lithosphere due to the emplacement of a thrust belt load. The lithosphere behaves as an elastic beam and flexes under the weight of the load forming the asymmetric shape characteristic of foreland basins (Beaumont, 1981; Jordan, 1981; Price, 1973). In peripheral foreland basins, such as the Persian Gulf (Fig. 6.5), that form by the closing of an ocean basin (Dickinson, 1976), the active subduction zone evolves into a collisional fold-thrust belt as one continent is partially subducted beneath another (A-type subduction, Bally, 1975). In retroarc foreland basins (Dickinson, 1976), such as forms adjacent to the subandean belt in South America (Fig. 6.5), intracontinental shortening initiates behind the volcanic arc but aborts as buoyant crust is forced under the upper plate (Coney, 1973). In either case, the total load consists of the tectonic load of the thrust sheets, derived sedimentary deposits and the emplacement of possible subsurface loads (Karner and Watts, 1983; Royden and Karner, 1984). In the case if peripheral foreland basins, the subsurface load, to a large extent, consists of the docking continental fragment, which acts as a load since it is replacing water that previously occupied the remnant ocean basin (Fig. 6.5; Stockmal and Beaumont, 1987; Stockmal et al., 1986).

The subsidence history of these basins reflects the rate of thrusting in the adjacent orogenic belt and sedimentation rate in the basin. The migration of thrust sheets into the basin results in an increasing rate of subsidence out across the basin producing a convex-up subsidence curve (Fig. 6.1). In addition, sediments derived from the thrust sheets after they are emplaced redistributes the load farther out across the basin forcing subsidence to migrate across the basin over time. Segmented subsidence curves (Fig. 6.1) may reflect discontinuous movement of the thrust belt. Erosion of the thrust load over long periods of time after thrusting ceases leads to removal of the thrust load and resultant rebound of the foreland basin (Heller et al., 1988). Subsidence, in this case, will cease and the basin will start to flexurally uplift in sync with the removal of the original load, redistributing sediment farther out across the foreland (see discussion in Chapter 7).

D. FOREARC BASINS

There has been relatively little quantitative study of the mechanisms of subsidence in forearc basins found in front of volcanic arc systems (Fig. 6.6). In part this reflects the difficulty in acquiring adequate stratigraphic data to accurately determine subsidence histories in this tectonic setting. Several mechanisms may play roles in generating forearc basin subsidence including:

(1) The development of a topographic basin between the topographic highs of the volcanic arc and accretionary wedge. In this case, all of the subsidence would result from the sediment load amplifying the previously existing topographic depression, thus there would be no "tectonic" subsidence. However, subsidence curves from forearc settings in which isostatic response due to sediment loading has been removed (Fig. 6.1), still show there to be a significant tectonic component to the subsidence history.

(2) Subsidence due to the isostatic response of emplacing a dense subducted plate beneath the forearc region. During initiation of a subduction zone, oceanic plate is consumed and underlies the forearc region, effectively doubling the lithosphere thickness along the convergent margin. Since oceanic lithosphere is denser than the asthenosphere that it replaces, the forearc region subsides. As the buoyancy of the subducted slab changes, either by changes in age of the

downgoing oceanic plate or by the subduction of thickened crust, such as oceanic plateaus, the density contrast between the subducted slab and asthenosphere changes leading to isostatic readjustment of the forearc basin. The resulting subsidence curve may show many vertical changes caused by these isostatic effects. Moxon and Graham (1987) plotted the subsidence history of the Mésozoic Great Valley Sequence in central California. The resulting subsidence curve contained abrupt changes in rate that they ascribed to a decrease in angle of subduction of the downgoing slab. The difficulty in applying this model as the sole or principle cause of forearc subsidence is that once subduction cease there is no longer a dense plate beneath the forearc region and the basin should isostatically rebound, uplifting and eroding the basinal sediments. In central California, where such a tectonic transition took place when subduction ceased in late Cenozoic time as a major transform fault, the San Andreas Fault, developed, no such rebound occurred. This suggests that this mechanism in itself does explain all of the subsidence history.

(3) A mechanism related to the subducted slab is subsidence caused by rapid cooling of a warm buoyant upper plate by conduction to a cooler underlying, subducted plate, as well as to the overlying atmosphere. In this case subsidence would be exponential over time and the magnitude of subsidence would be similar to other thermally-driven basins (such as passive margins), but the rate of subsidence would be much faster than a cooling plate by itself. This mechanism would only apply to situations where hot, young oceanic crust which accrete along convergent margins become, soon thereafter, sites of forearc subsidence. This situation, albeit unusual, might explain both the modern Japan and Eocene Oregon examples (Fig. 6.1).

(4) Another possible mechanism for forearc subsidence is flexural back-tilting of the basin floor under the topographic load of the volcanic arc and by the underthrusting and build up of the accretionary wedge (in a sense an upward-directed, or buoyant, load). Bond et al. (1988) have each suggested that the volcanic arc may act as a part of the load causing flexural subsidence in the North Aleutian basin, a back-arc settings behind the Aleutian arc. It may be possible that similar loading by the arc may lead to subsidence in the forearc region as well. A fundamental question with this model is whether or not active volcanic arcs are locally

isostatically compensated and, therefore, do not exert a load for long wave-length flexure of the lithosphere. In contrast, sedimentary underplating and uplift of the accretionary subduction complex in front of the forearc region generates an upward-directed, or buoyant, load that may cause back-tilting of the forearc region in flexural response. In the Arica and Iquique forearc basins (Fig. 6.7), off the coast of Chile, progressive back tilting of forearc sedimentary packages over time (Coulbourn and Moberly, 1977) suggests a progressive loading origin, either by the volcanic arc, the accretionary subduction complex or both.

Limited data suggests that subsidence histories of various forearc basins seem to be quite different (Fig. 6.1), suggesting that multiple mechanisms may be at work.

E. INTRACONTINENTAL BASINS

Large circular cratonic basins, such as the Michigan Basin (Figs. 6.8, 6.9, and 6.10), are relatively rare on earth, but have caught the eye of subsidence modelers. Subsidence of these basins is rather slow in comparison to other basin types and to the subsidence of the aging ocean floor (Fig. 6.1). Subsidence lasts many tens to hundreds of million years with only one or two kilometers of tectonic subsidence over this time. Subsidence in some of these basins appears to be stepped (Fig. 6.11). Nonetheless, most models of basin formation call on thermal subsidence mechanisms (e.g., Haxby et al., 1976; Heidlauf et al., 1986; Klein and Hsui, 1987; Nunn and Sleep, 1984; Nunn et al., 1984). In the Haxby et al. (1976) model (Fig. 6.12), a thermal event pierces the mantle lid, warms up the base of the crust causing it to convert to a higher-density phase. Once the lithosphere cools, the dense lower crust acts as a sinker which flexurally pulls down the surrounding lithosphere forming a bull's eye-shaped basin.

Other mechanisms of formation for intracontinental basins have been suggested, including: transform-related basins associated with continental collision and escape (Kluth and Coney, 1981; Molnar and Tapponier, 1975), and failed rifts (Dickinson, 1976). However, the exact nature of the mechanisms of formation for these basins still proves to be difficult to interpret.

F. SUMMARY

The basic mechanisms of subsidence (i.e. local isostasy, flexure and thermal) can be applied, in many cases in a relatively straight forward way, to most major tectonic settings. For the most part, these applications are rather broad, they describe the overall subsidence style but do not explain details of the basin history. The major hinderance to a more complete understanding of basin evolution is the lack of data available for generating well constrained subsidence histories in several tectonic settings. In addition, such analyses need to be done in two and three dimensions, and not just the one-dimensional analyses done here. Certainly there is much to be learned about lithospheric mechanics and sedimentary basin development, as well as sedimentary and tectonic interactions once more complete data sets become available.

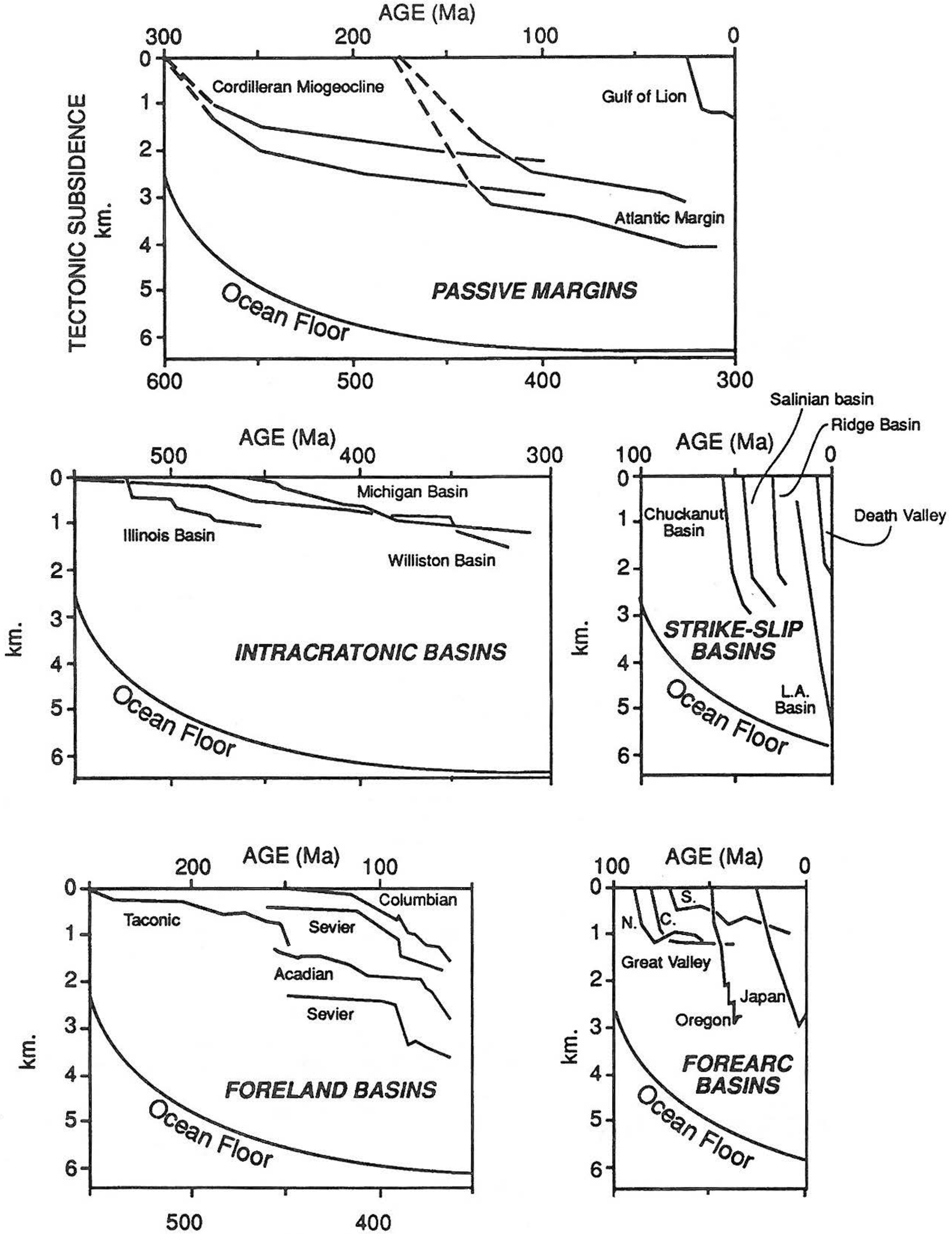

Figure 6.1 Representative tectonic subsidence histories of basins from different tectonic settings (Heller et al., in prep.).

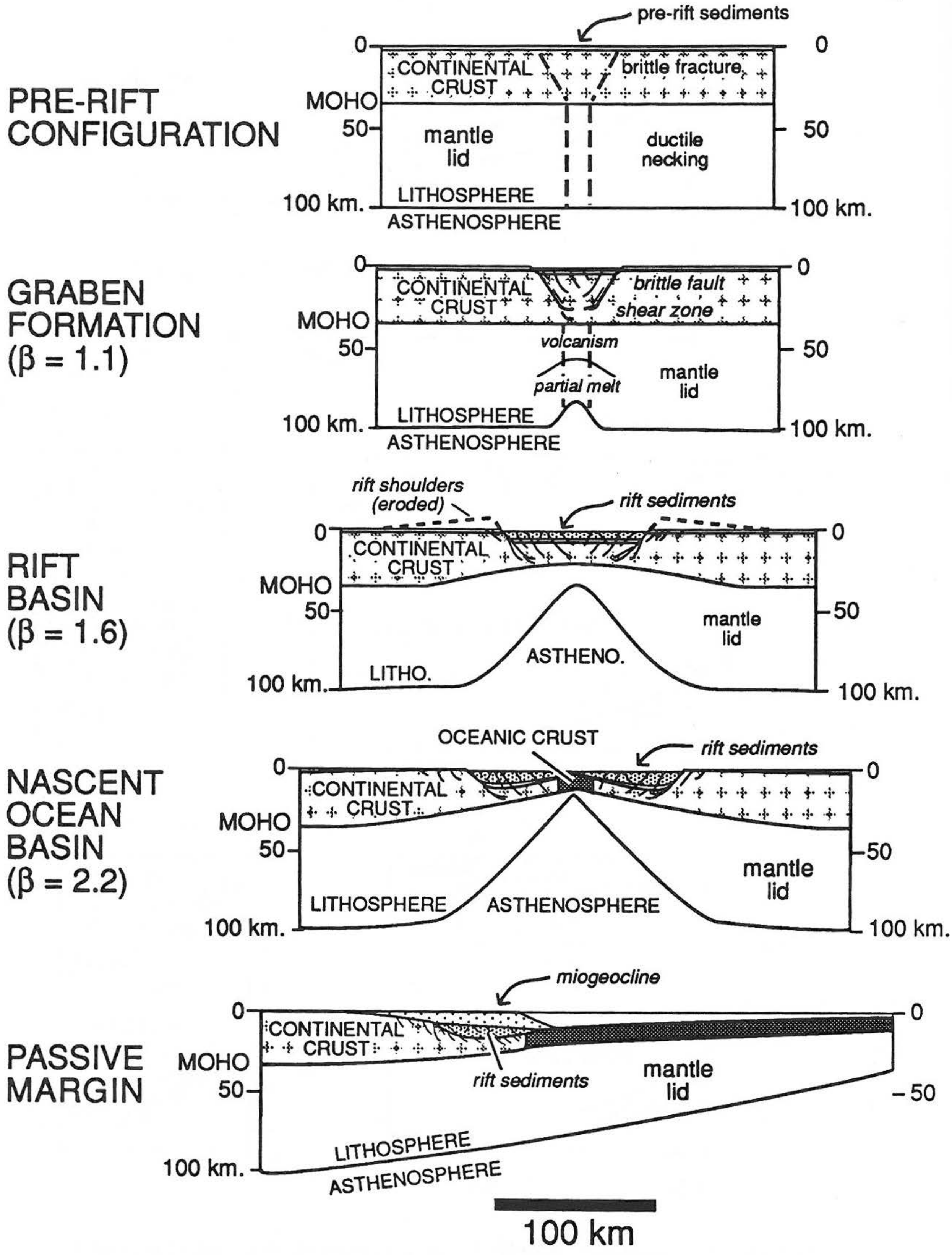

Figure 6.2 Diagrammatic evolution of rift basins and passive margins. Modified from Salveson (1978).

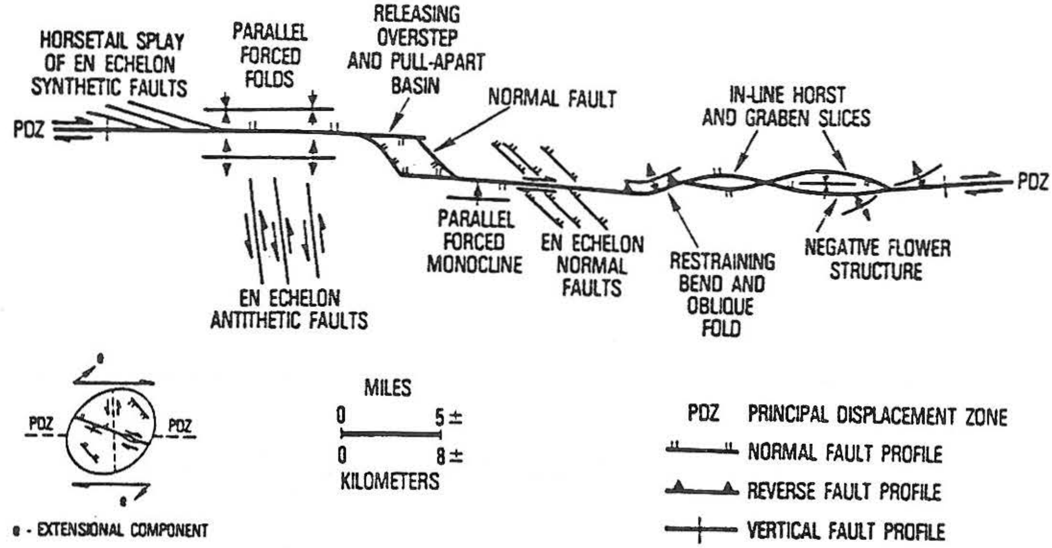

Figure 6.3A Assemblage of structures associated with right-slip divergent wrench fault. From Harding et al. (1985).

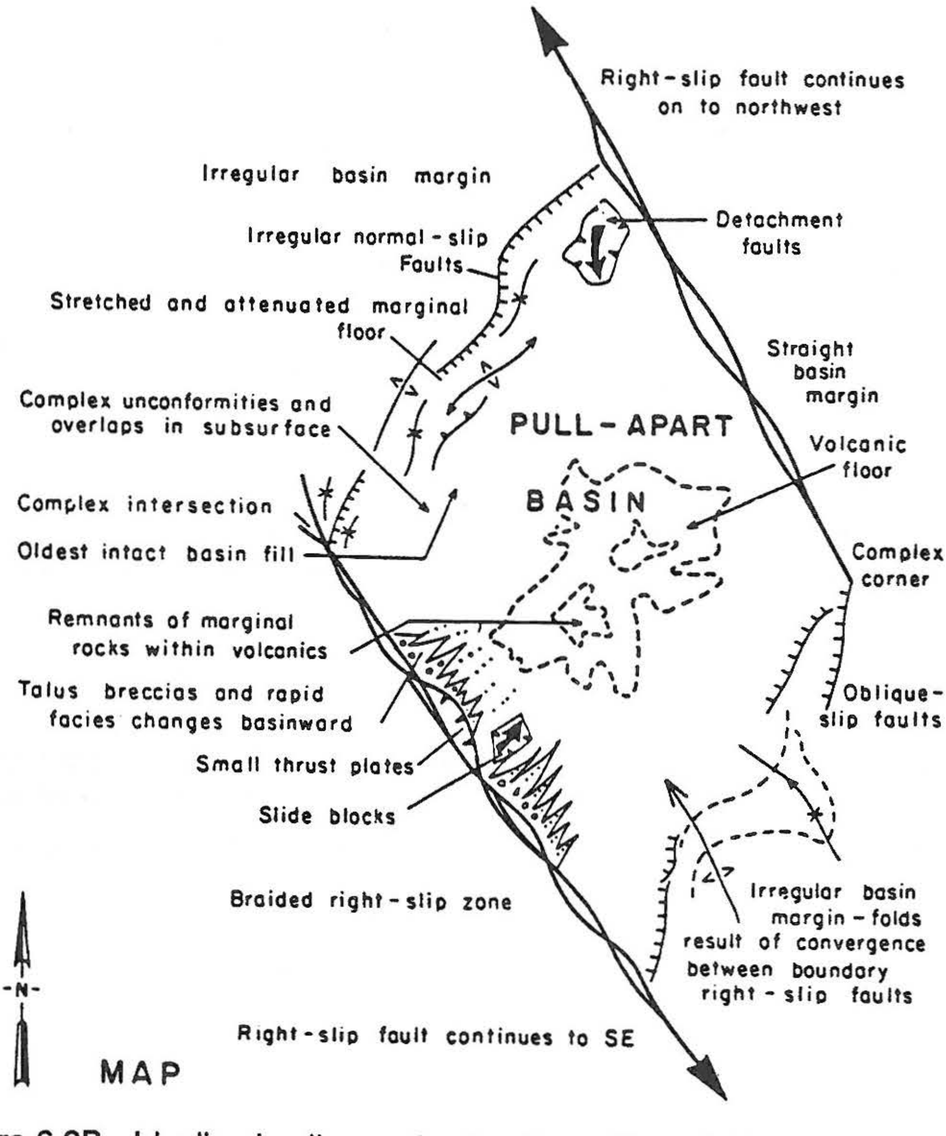

Figure 6.3B Idealized pull-apart basin. From Crowell (1974).

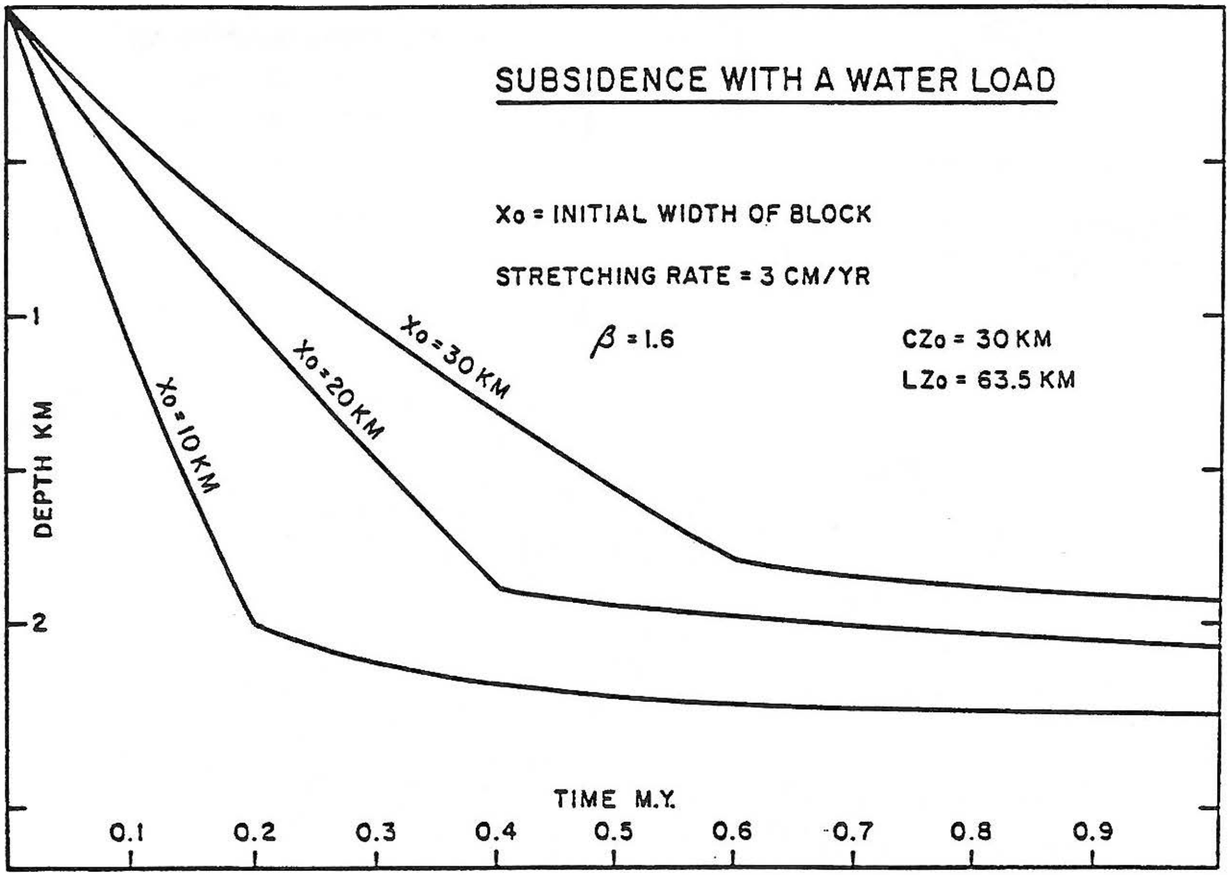

Figure 6.4 Idealized subsidence history of narrow rift basins (from Pitman and Andrews, 1985). Shows subsidence histories for basins of three different initial block widths (X_0). Each block is stretched at 3 cm/yr until $\beta = 1.6$. CZ_0 and LZ_0 are initial crustal and lithospheric thicknesses, respectively.

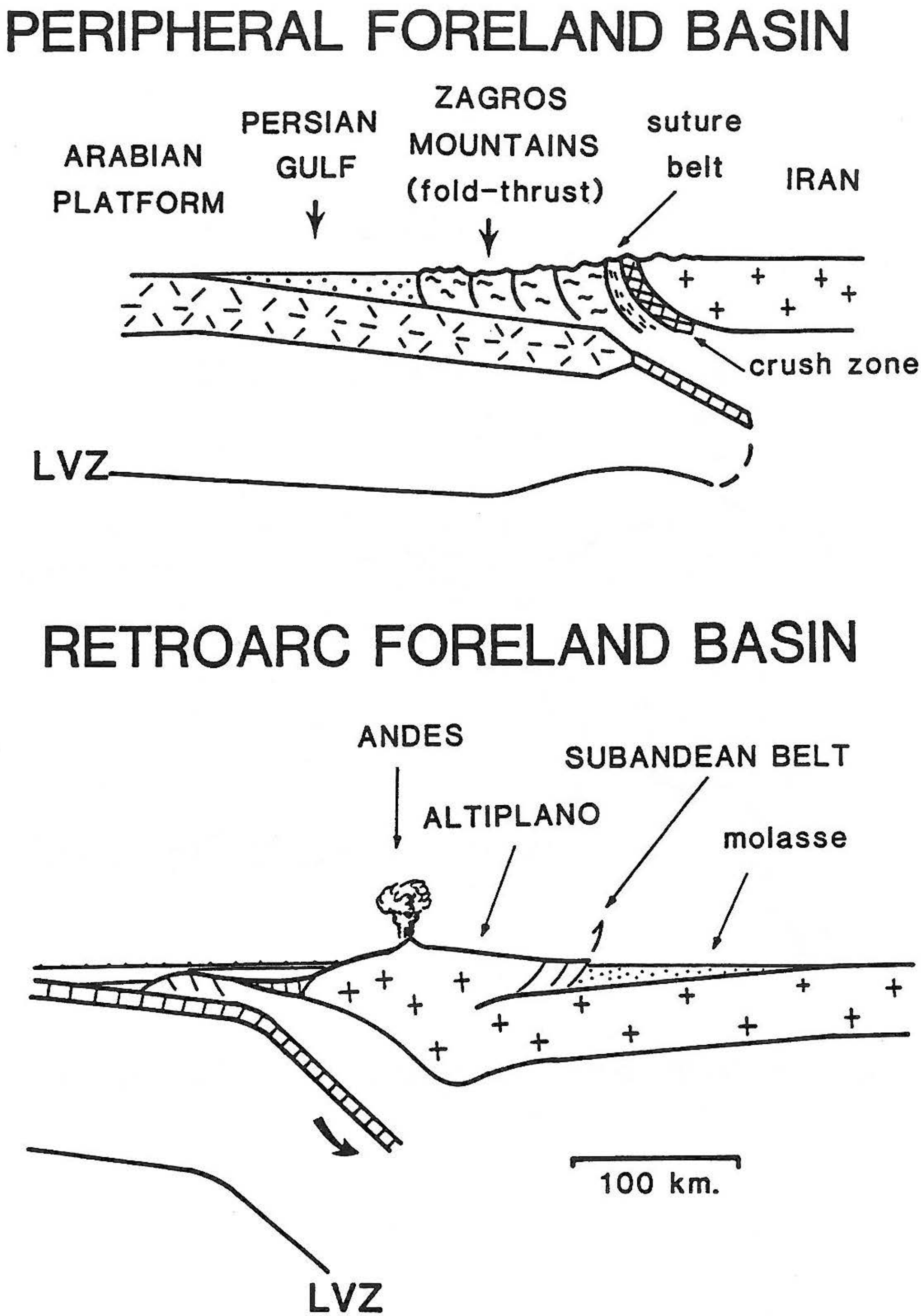

Figure 6.5 Tectonic setting of peripheral and retroarc foreland basins. Based on unpublished figures from W.R. Dickinson (1982).

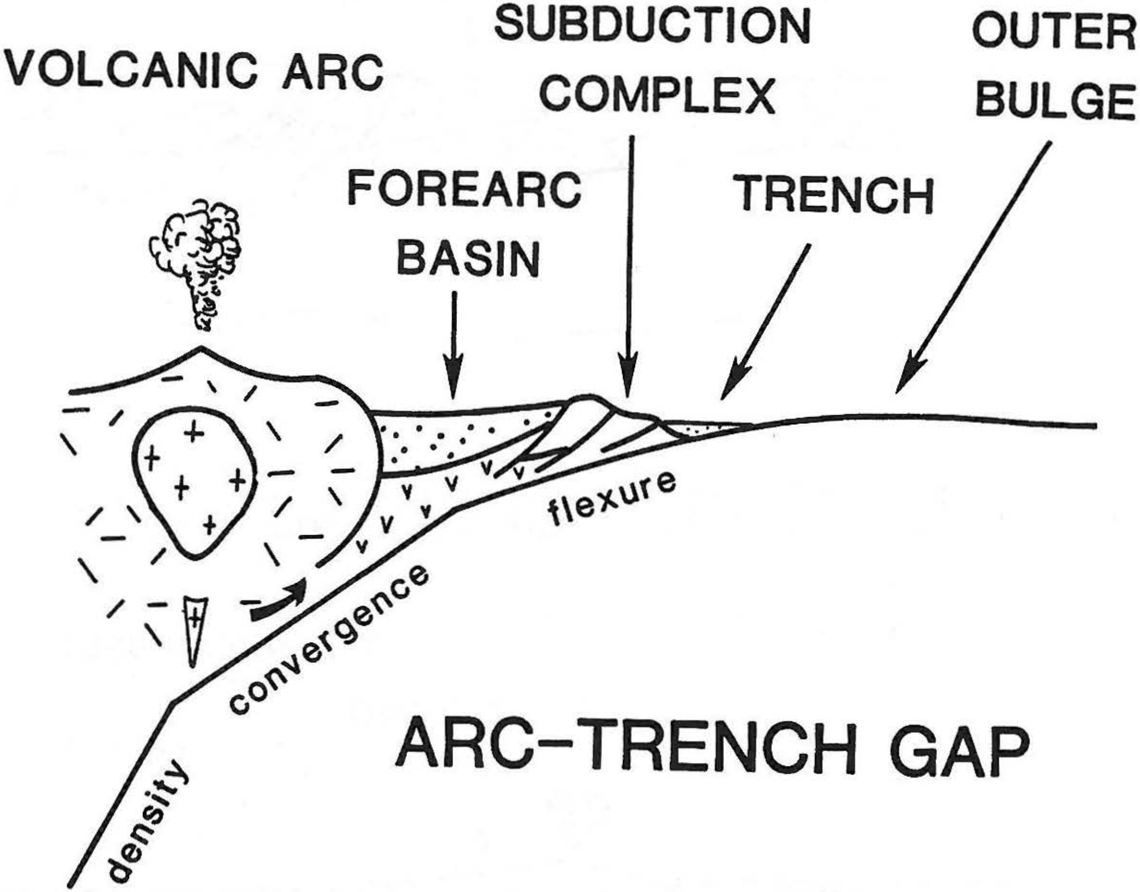

Figure 6.6 Tectonic setting of arc-trench gap and forearc basins. Based on the Sunda arc (Curry et al., 1977).

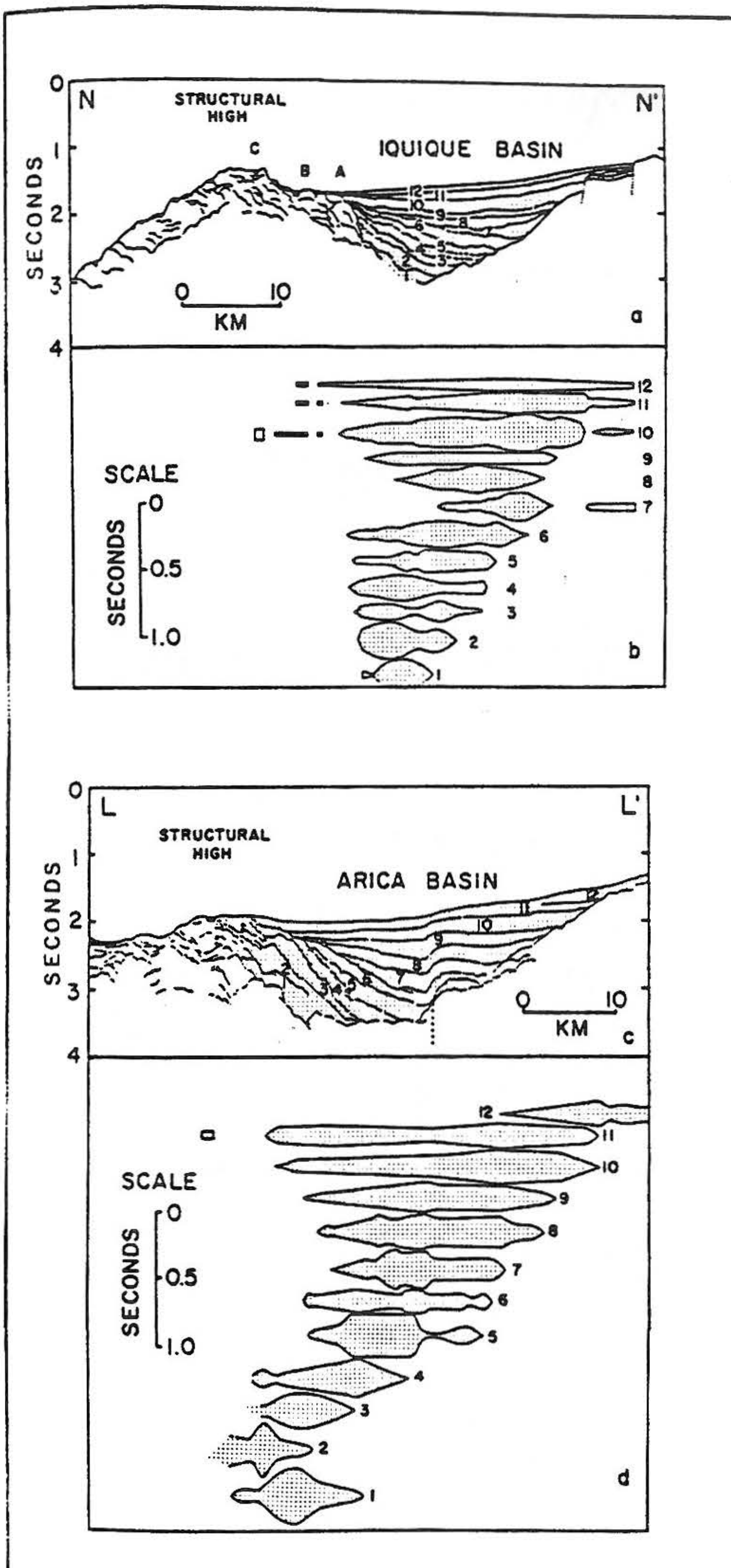

Figure 6.7 Profiles of Iquique and Arica Basins, South America, including plot of two-way travel time between reflectors versus lateral distance. From Coulbourn and Moberly (1977).

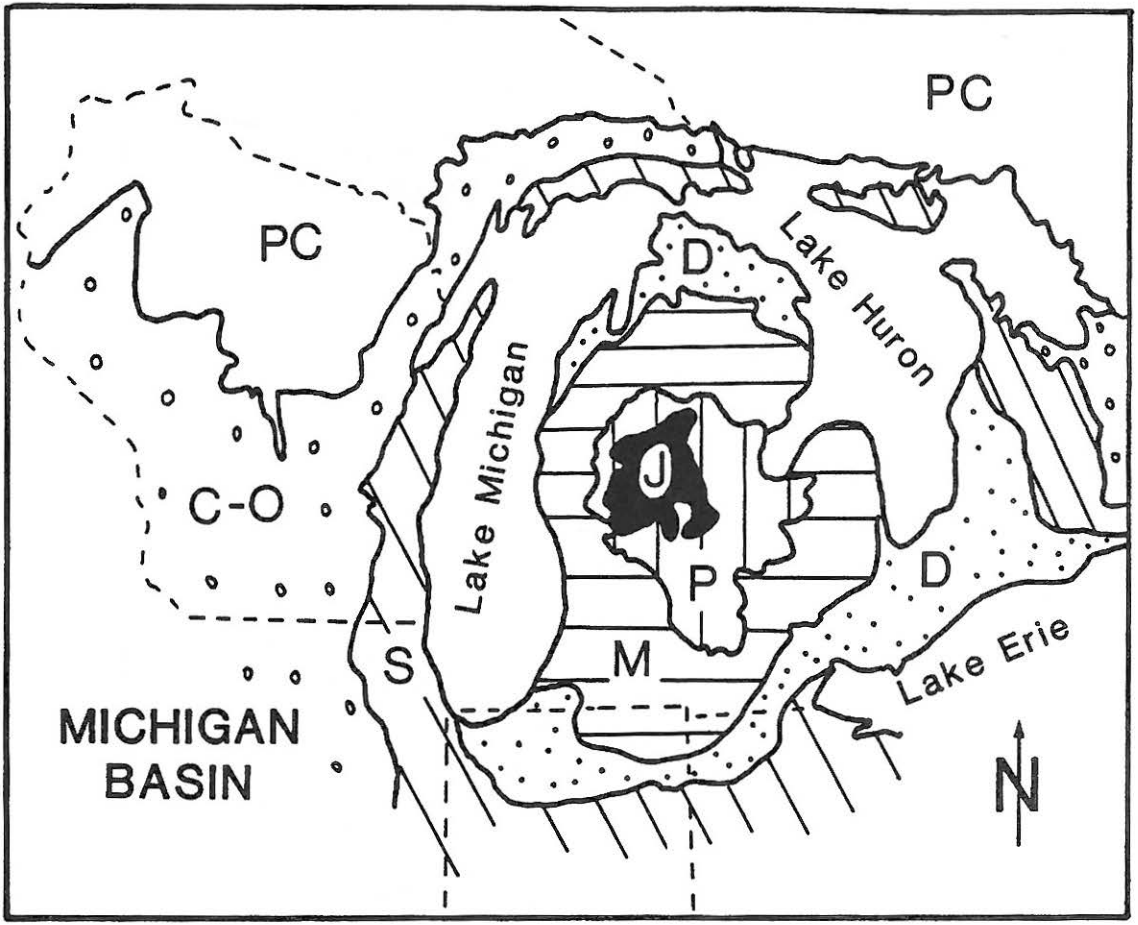

Figure 6.8 Simplified geologic map of the Michigan Basin. Modified from Nunn et al. (1984a).

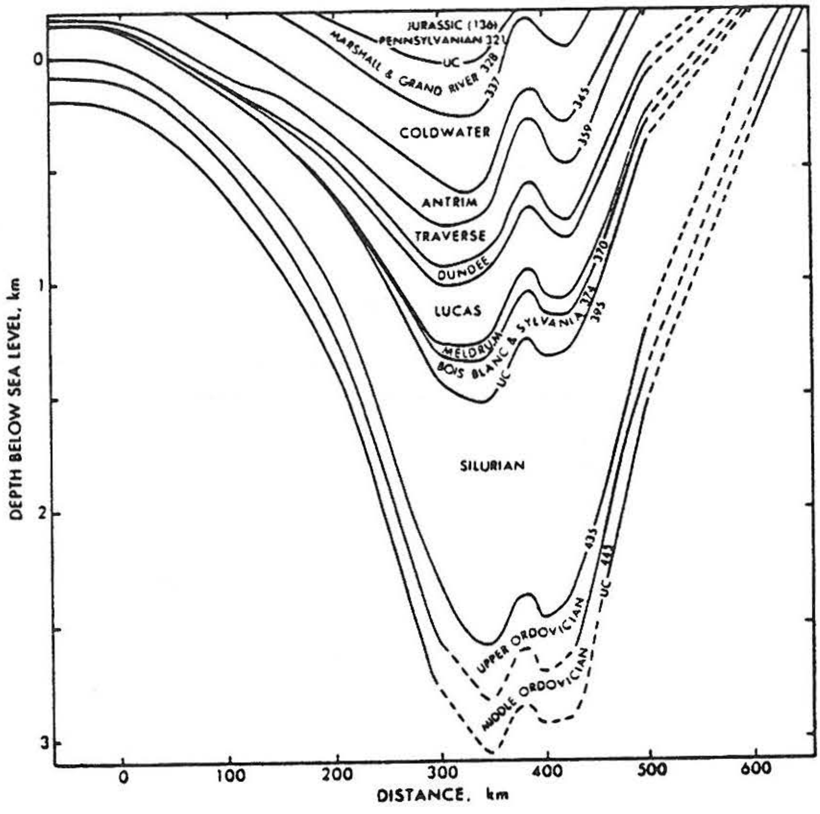

Figure 6.9 Ordovician to Jurassic cross-section of the Michigan Basin. From Nunn et al. (1984a).

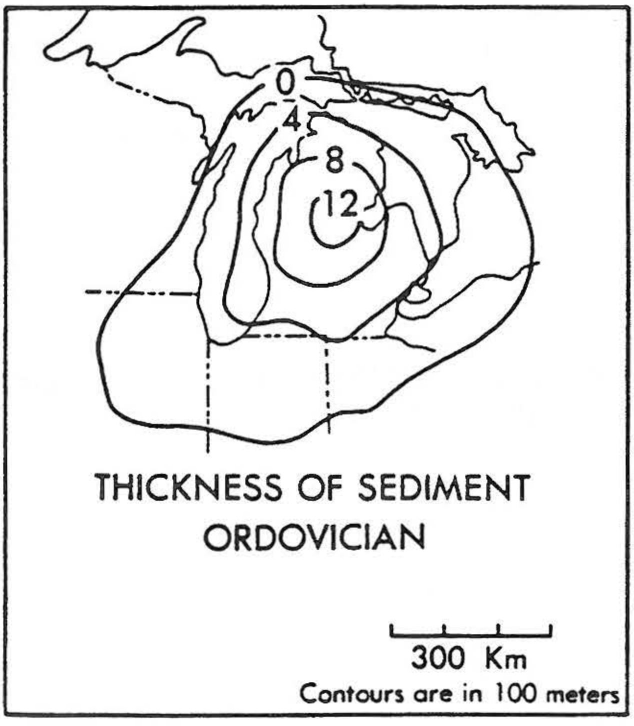

Figure 6.10 Ordovician isopach map of the Michigan Basin. From Nunn et al. (1984a).

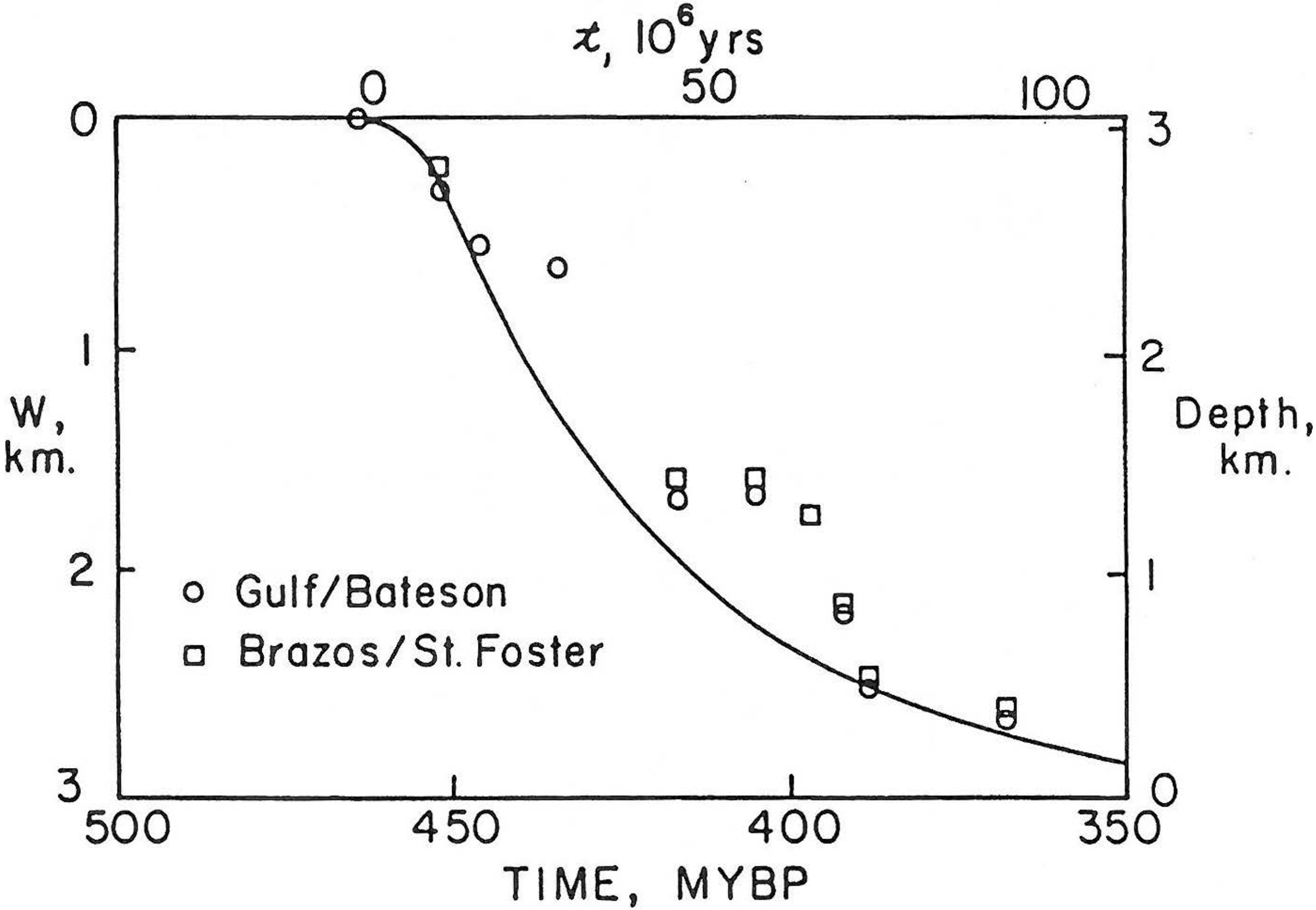

Figure 6.11 Total subsidence history of the Michigan Basin during middle Paleozoic time. From Haxby et al. (1976).

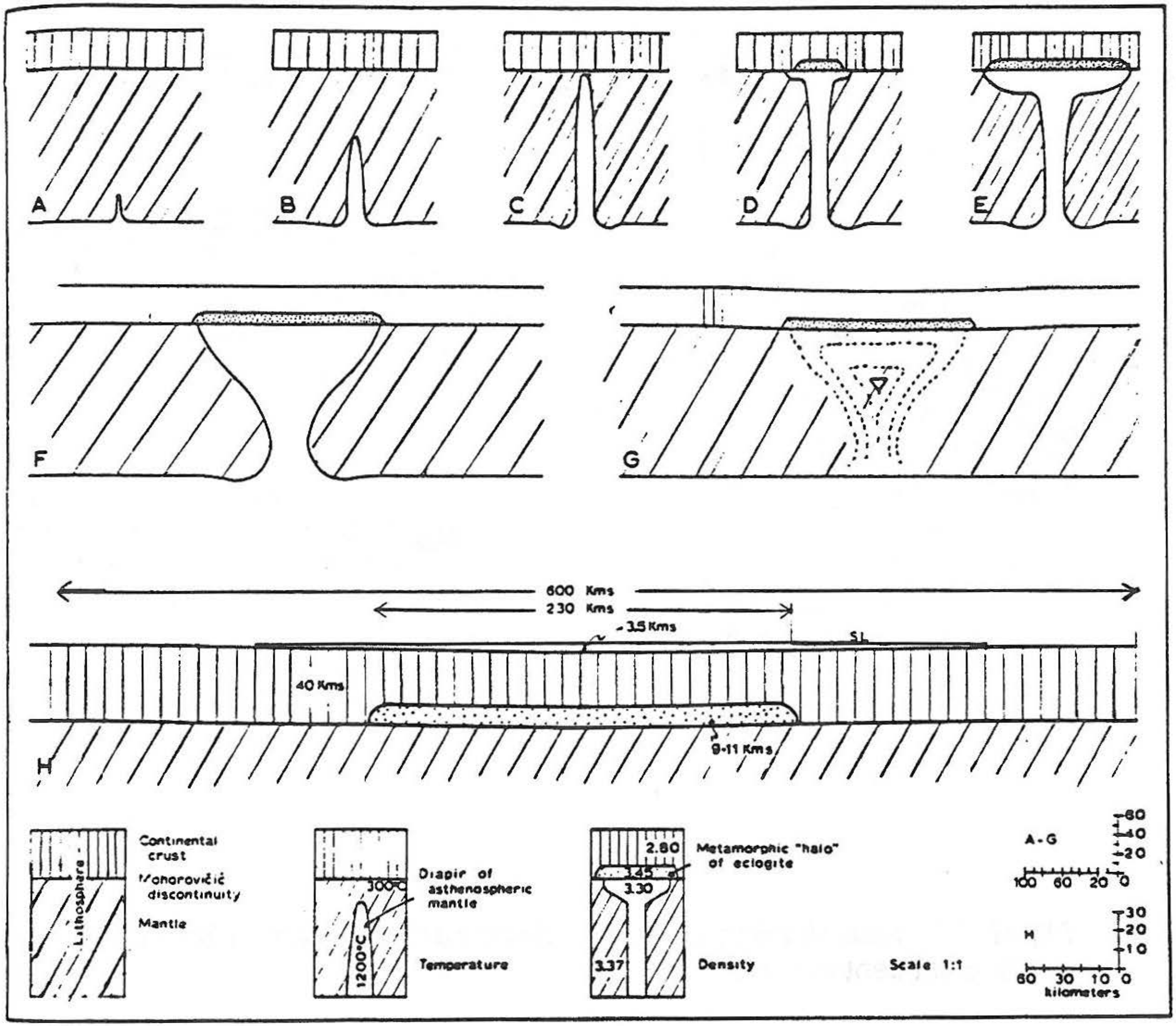

Figure 6.12 Insitu densification model for formation of the Michigan Basin. From Haxby et al. (1976).