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Notes

Plate tectonics and basin subsidence history

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ABSTRACT

Tectonic setting exerts first-order control on basin formation as reflected in basin subsidence history. While our approach ignores the effects of flexural loading and eustatic sea-level change, consistency of backstripped subsidence histories (i.e., with local loading effects of sediment removed) suggests consistent tectonic driving mechanisms in each tectonic setting, with the possible exception of forearc basins.

Based on published subsidence curves and open-file stratigraphic data, we show the subsidence characteristics of passive margins, strike-slip basins, intracontinental basins, foreland basins, and forearc basins. Passive margin subsidence is characterized by two stages, rapid initial, synrift subsidence and slow post-rift thermal subsidence, with increasing subsidence rates toward the adjacent ocean basin. Subsidence of intracontinental basins is similar in magnitude to that seen in passive margin settings, but the former is generally slower, longer lived, and lacks initial subsidence. Long-lived subsidence for many intracontinental basins is consistent with cooling following thermal perturbation of thick lithosphere found beneath old parts of continents. Basins associated with strike-slip faults are usually short lived with very rapid subsidence. Changes in local stress regimes as strike-slip faults evolve, and migrate over time, coupled with three-dimensional heat loss in these small basins likely explain this subsidence pattern. Foreland basin subsidence rates reflect the flexural response to episodic thrust loading. Resultant subsidence curves are punctuated by convex-up (accelerating) segments. Forearc basins have the least consistent subsidence patterns. Subsidence histories of these basins are complex and may

reflect multiple driving mechanisms of subsidence in forearc settings.

Second-order deviations in subsidence suggest reactivation or superimposed tectonic events in many basin settings. The effects of eustatic sea-level change may also explain some deviations in curves. For many of these settings, subsidence histories are sufficiently distinctive to be used to help determine tectonic setting of ancient basin deposits.

Keywords: subsidence analysis, passive margins, intracontinental basins, foreland basins, strike-slip basins, forearc basins.

INTRODUCTION

Sedimentary basins reflect prolonged subsidence of Earth's surface, due to large-scale tectonic processes operating between and within plates (Kuszniir and Ziegler, 1992). To the degree that tectonic processes are reflected in subsidence history, basins in similar tectonic settings should show similar patterns of subsidence. Subsidence histories are taken to reflect isostatic adjustment to lithospheric processes, such as thermal events, thickness changes, and loading history. Therefore, subsidence history provides insight into basin-forming mechanisms. Differences in subsidence histories between basins may reflect how fundamental driving mechanisms vary as well as secondary influences, such as sea-level change and sediment loading. By comparing subsidence curves between different basins of similar tectonic setting, it is possible to determine consistency and/or differences in the processes that drive subsidence.

Dickinson (1976) and Angevine et al. (1990) compiled subsidence histories of basins in order to discriminate between subsidence styles in various tectonic settings. Since those studies, more data have become available that may be used to more fully define subsidence patterns as a function of tectonic setting. In this paper, we use some of these data to demonstrate the styles of subsidence in various plate tectonic settings, emphasizing those settings where the modes

of subsidence are still poorly understood. Our objective is to show that these histories can be used as templates allowing tectonic interpretations based on subsidence patterns.

SUBSIDENCE ANALYSIS

Subsidence analysis yields a graphic representation of the vertical movement of a stratigraphic horizon, with respect to a datum in a sedimentary basin. It tracks the subsidence and uplift history at that location since the horizon was deposited (van Hinte, 1978). Data needed to reconstruct subsidence history include stratigraphic thickness, lithology, estimate of paleowater depths, and age control. Subsidence analysis begins with a plot of sediment accumulation through time using the present-day thickness of each dated stratigraphic unit. Second, the effects of compaction are included based on the assumption that porosity lost is mostly caused by mechanical compaction. Third, since sea level is used as the datum for subsidence analysis, paleobathymetry corrections are needed to correct the seafloor position to this datum. The resulting curve reflects the total subsidence history (van Hinte, 1978) including the contribution of tectonic loads, sediment loads, and sea-level changes. Of these, the local isostatic effects of sediment loading can be removed by "backstripping" (Steckler and Watts, 1978). The resulting subsidence curve, referred to as "tectonic subsidence," shows the idealized subsidence history of a basin that would have existed if only water, and no sediment, filled the subsiding hole. Tectonic subsidence history reflects basin subsidence due to factors other than sediment deposition and attendant isostatic adjustment and compaction. More importantly, it provides a way of normalizing subsidence in different basins that have undergone very different sedimentation histories.

It is important to be aware of some limitations to this analysis. These come from the inaccuracy of data used to reconstruct history and from the assumptions built into the method. In particular, age control and water depth often hamper

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subsidence analysis. Water-depth estimates are often difficult to determine because of a paucity of unique depth indicators in sedimentary rocks. However, the impact of poorly constrained water depth can be reduced by analyzing sections that are mostly composed of shallow marine deposits, thus reducing the absolute magnitude of water-depth uncertainties. In addition, any uncertainties due to water-depth assignments are reduced by working with relatively thick stratigraphic successions and relatively shallow water depths. For our compilation, we have chosen to use a simplified water-depth scale (Angevine et al., 1990), assigned based on fossil and/or lithofacies evidence: nonmarine is here considered to be 50 m above sea level, inner shelf is 50 m, outer shelf is 150 m, and upper slope is 350 m. Non-marine water depths are typically from coastal plain deposits associated with shorelines. For the most part, errors in water-depth assignment are relatively small compared to the magnitudes of the curves. However, significant changes in water depth occurring within the large uncertainties inherent in bathyal-abyssal water depths may leave important tectonic signals undetected (Dickinson et al., 1987) and so are not included here. Age control is another potential source of inaccuracy. The number and resolution of age assignments vary widely for different sedimentary successions. We have focused on sections that have at least four to five dated stratigraphic horizons identified as points for subsidence analysis. For this compilation we accept the time scales used by the original authors. The possibility of significant hiatuses can also

lead to error. In order to reduce the impact of this issue, we limit our study to those sections where major unconformities, where identified by the original authors, are few. We compensate for changes in stratigraphic thickness through time due to compaction following the standard approach outlined by van Hinte (1978) and Allen and Allen (2005). For simplicity, we used the exponential porosity versus depth relationships from Sclater and Christie (1980), and generalized lithologies to sandstone, shale, or limestone. Subsidence is calculated with respect to sea level. Since no universally accepted quantified curve of sea level exists, we have simply chosen to ignore sea-level changes and assume that they cause only relatively low-magnitude variations in our calculated subsidence. Even the magnitude of long-term sea-level change is poorly constrained, but is likely small (<200 m) relative to the magnitude of subsidence seen in these curves (Haq et al., 1987; Harrison, 1990). As such, we assume that the sea-level datum has not changed over time. Our approach to backstripping utilizes the one-dimensional local isostatic method of Steckler and Watts (1978). However, this approach does imply that significant variations in sediment loading near the analyzed site in the basin do not cause subsidence. For broad basins, where basin thicknesses do not vary greatly over short distances, this is a reasonable assumption (Angevine et al., 1990). This becomes more important for relatively small basins and/or those formed over relatively rigid lithosphere. This error is reduced by the fact that we are looking at comparing the overall

shape of resultant curves in similar tectonic settings and in basins of similar size.

All subsidence curves presented here are tectonic subsidence curves. Subsidence histories were only collected from sites where the plate tectonic setting, as reported in the literature, is well known (Fig. 1). It is possible that future work will lead to a reinterpretation of tectonic setting for some of these curves. We document both the location and the original author used in our compilation so that these redefined cases can be clearly identified. Subsidence curves are grouped together by tectonic setting and plotted at the same scale to facilitate comparisons (Figs. 2–7). For clarity, several simplifying approaches are used to improve our comparison of curves. First, no uncertainties in age and paleobathymetry are shown on the graphs. Second, major unconformities, if present, are shown by flat horizontal lines on the subsidence curves, and small unconformities are not shown.

RESULTS

Our compilation is limited to sources that provide quantitative data that include stratigraphic thickness, age assignments, proxy water-depth data, and where the tectonic setting is well established. In addition, an individual basin may undergo different phases of subsidence as tectonic settings change through time. We have tried to isolate those phases, or megasequences (Allen and Allen, 2005), during which the specific tectonic setting of interest occurs. Figure 1 locates the sites and tectonic setting for

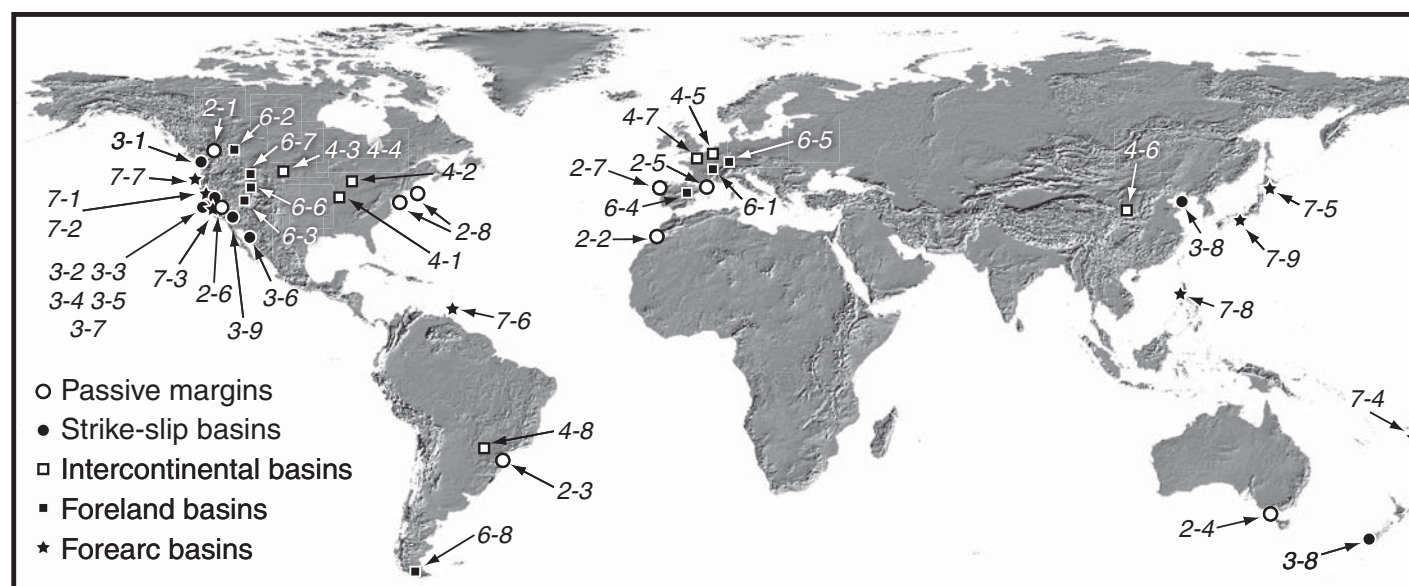


Figure 1. Locations of subsidence data by tectonic settings. Numbers refer to specific subsidence curves. First value is figure number, and second value is curve number in that figure (e.g., 2–3 = Figure 2, curve 3).

data compiled in this study. Figures 2–7 show the results of subsidence analysis as a function of plate tectonic settings. On each set of subsidence curves we include, for comparison, a reference curve that parallels best-fit thermal subsidence of the seafloor assuming a semi-infinite half-space model from Stein and Stein (1992). All curves are corrected for compaction and backstripped assuming local isostasy.

Passive Margins (Fig. 2)

Subsidence following continental rifting and breakup leads to asymmetric subsidence and foundering of continental margins (Steckler and Watts, 1978). As a result, the amount of subsidence increases seaward of the hinge zone. All subsidence curves show an initial phase of rapid subsidence followed by a phase in which subsidence rates are reduced (e.g., Watts and Ryan, 1976; Steckler and Watts, 1978) and mimic the age-depth curve of the seafloor. Some margins demonstrate an abrupt change in subsidence rates between these phases. However, this abruptness may reflect a poorly constrained history of early subsidence in some cases. Initial subsidence deposits are often coarse nonmarine deposits that are notoriously difficult to biostratigraphically date. In addition, in modern settings, the initial subsidence deposits are the deepest and, thus, less frequently penetrated parts of the sections. As a result there tend to be few age constraints to delimit the early subsidence history and the transition from rapid to slower subsidence. Nonetheless, some well-constrained curves, such as the one shown from the Gulf of Lion (Steckler and Watts, 1980), indicate that abrupt changes can be real. Subsidence in passive-margin settings typically continues for more than 150 m.y. Maximum subsidence (Fig. 2) varies up to 4 km, in part depending on distance seaward of the hinge zone (i.e., the landward limit of extension).

Passive margin formation and subsidence mechanisms have been much studied following the breakthrough work of Watts and Ryan (1976) and Steckler and Watts (1978). Rift basins develop early during continental breakup followed by passive margin subsidence once breakup is complete. Not all rifts go to completion, and many “failed rifts” can be found (cf. Allen and Allen, 2005, their Fig. 9.11). Theoretical and analytical studies suggest that tectonic subsidence can be divided into an initial “synrift” phase that primarily reflects isostatic response to extension and thinning of continental crust, followed by a “post-rift” phase driven by thermal reequilibration as the lithosphere cools and thickens back to equilibrium. Synrift stretching and thinning by factors of less than 2 are com-

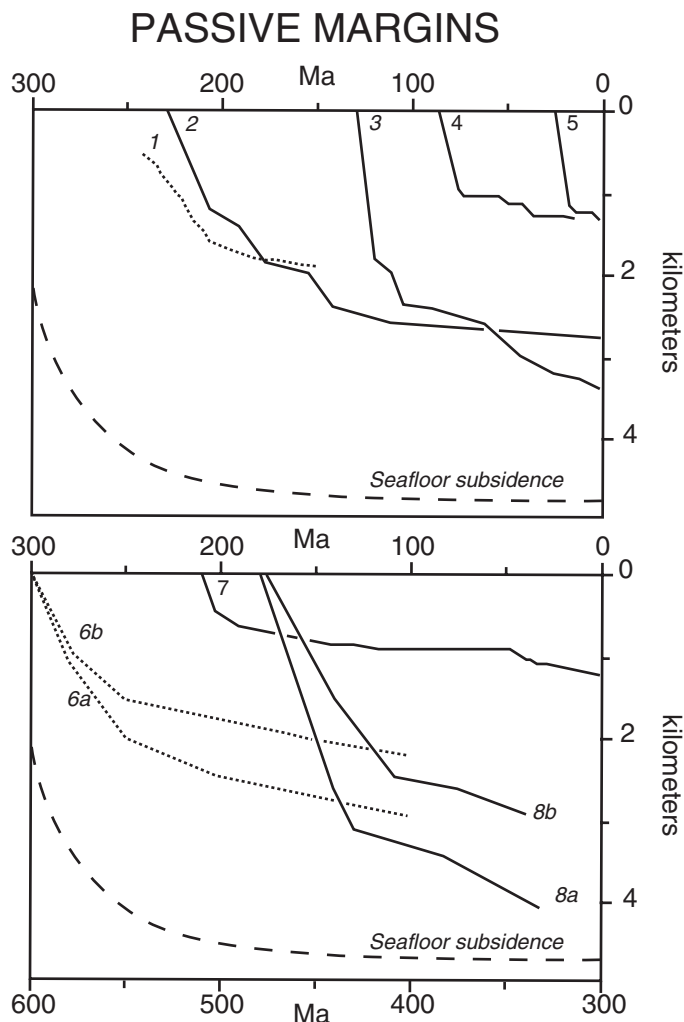


Figure 2. Tectonic subsidence curves for passive margin settings. Locations shown on Figure 1. Solid curves correspond to time scale at top of graph and dotted lines to time scale at bottom of graph. Thermal decay curve (dashed) for subsidence of cooling seafloor (Stein and Stein, 1992), minus (i.e., shallowed) 500 m, is shown for comparison. 1—Paleozoic Miogeocline, southern Canadian Rocky Mountains (Bond and Kominz, 1984); 2—Moroccan Basin (Ellouz et al., 2003); 3—Campos Basin (Mohriak et al., 1987); 4—Gippsland Basin (Falvey and Mutter, 1981; P. Yin, 1985, personal commun.); 5—Gulf of Lion (Benedicto et al., 1996); 6—U.S. Cordilleran Miogeocline (Bissell, 1974; Armin and Mayer, 1983; Devlin et al., 1986; Devlin and Bond, 1988); 7—Lusitanian Basin (Stapel et al., 1996); 8—U.S. Atlantic margin (Steckler and Watts, 1978; Swift et al., 1987).

mon in rift basins (e.g., Hendrie et al., 1994; Kuszniir et al., 1996a, 1996b; Roberts et al., 1995; Swift et al., 1987) and variable along individual passive margins. In general, stretching factors increase seaward to the point of continental rupture and ocean crust formation. In addition, local variability in subsidence can reflect local structure and thinning as well as superimposed effects

(King and Ellis, 1990; Nadin and Kuszniir, 1995). Various mechanical models have been proposed to explain details of subsidence curves in this setting. Such models consider how extensional strain is partitioned through the lithosphere (e.g., pure shear versus simple shear and depth-dependent stretching), character (e.g., symmetric versus asymmetric and volcanic versus nonvolcanic

margins), timing, and rate of heat loss during and following continental breakup (e.g., Bott, 1980; Jarvis and McKenzie, 1980; Turcotte, 1980; Watts, 1981; Wernicke et al., 1982; Cochran, 1983; Nadon and Issler, 1997), as well as superimposed tectonic events (e.g., Dore and Stewart, 2002; Nielsen et al., 2002). Variations in these factors may explain differences in magnitude of curves shown.

Strike-Slip Basins (Fig. 3)

Basins related to strike-slip faults include a variety of basin types that result from a combination of transform fault movement that may include either elements of crustal extension or shortening (Christie-Blick and Biddle, 1985). Strike-slip basins show a large variety in basin size and geometry. However, they are typically narrow and smaller than those produced by regional extension or shortening. Different types of basins can form in strike-slip settings, from simple pull-apart basins developed along fault oversteps to more complex basin forms in zones of transtension and transpression (e.g., May et al., 1993; Sutherland and Melhuish, 2000). Fault geometry and related fault-mechanical process are the critical controls for the development of strike-slip basins as demonstrated by different authors (e.g., Crowell, 1974; Mann et al., 1983; Ingersoll, 1988). Regardless of specific basin shapes and locations, all curves generated for this setting are characterized by rapid and short-lived (typically <10 m.y.) subsidence.

Subsidence in this setting is dictated by the spatial configuration of the various scales of strike-slip fault systems within the basin, as well as the history of displacement and attendant heat loss (Sawyer et al., 1987; Chen and Nábelek, 1988). Depending on local patterns of deformation, subsidence curves in strike-slip basins may be episodic and end abruptly (e.g., Crowell, 1974; Mann et al., 1982; May et al., 1993). Tectonic subsidence typically exceeds 2 km and reaches 4 km in exceptional cases (e.g., the Los Angeles basin, Fig. 3, line 5). The magnitude and concave-up shape of these curves are similar to those from passive margin settings (Fig. 1), although the subsidence rates are much faster. The fact that most strike-slip basins are short lived probably reflects the evolution and frequent change in position of the master strike-slip faults (Sylvester, 1988; Cloetingh et al., 1996; Storti et al., 2003; Allen and Allen, 2005; Waldron, 2005).

Stretching models have been applied to strike-slip basins; however, compensation is made for the small space and time scales associated with these basins, such as finite rifting times, accentuated lateral heat flow, and depth-dependent

extension (Cochran, 1983). Heat flow increases due to lithospheric thinning. Theoretical and field studies suggest that heat is lost rapidly during the extension process, in part, by lateral conduction (Cochran, 1983; Pitman and Andrews, 1985). The short-lived tails seen at the ends of most curves may result from cooling of the small remaining thermal anomaly once the fault ceased to be active (Pitman and Andrews, 1985). The result is that there is very little subsidence that continues once extension stops (Nilsen and McLaughlin, 1985). In some cases, the absence of evidence for post-rift thermal subsidence may also be a result of subsequent deformation of the basin (Christie-Blick and Biddle, 1985).

Intracontinental Basins (Fig. 4)

Intracontinental, or intracratonic, basins are large basins formed on an old continental lithosphere away from any known active tectonic margin (Dickinson, 1976). These basins are typically quite large (>150,000 km² in area), have relatively slow, long-lived subsidence (typically >200 m.y. in duration), but in most compiled cases tectonic subsidence is less than 2 km. Cross-sectional geometries of the North American examples are approximately symmetric (Illinois, Michigan, and Williston Basins); whereas, others are not.

Subsidence curves of intracontinental basins in Figure 4 are approximately exponential in shape, similar to passive margins, but most lack a rapid initial subsidence phase. Overall, subsidence curves follow the shape and magnitude of seafloor subsidence, but with longer decay constants. Such a comparison has led some (Haxby et al., 1976; Sleep and Sloss, 1980; Cercione, 1984; Nunn and Sleep, 1984; Nunn et al., 1984; Howell and van der Pluijm, 1999; Kominz et al., 2001) to suggest a thermal decay origin for at least some of these basins.

A comparison of intracontinental subsidence curves to simple thermal subsidence models (Fig. 5) indicates broad consistency. To model thermal subsidence, we use McKenzie's (1978) simple-stretching model, but only calculate post-rift subsidence resulting from lithosphere reequilibration following thinning. Stretching (thinning) factors in this case only reflect thinning of mantle lithosphere due to thermal perturbation and not necessarily extension. Stretching factors ranging from 1.1 to 1.5 and equilibrium lithosphere thickness of 125 and 200 km are shown. Notice that thicker lithosphere has a longer decay constant to reach thermal equilibrium. Thermal decay constants increase as the square of lithosphere thickness. As a result, it is not too surprising that thick lithosphere, which tends to exist beneath the

oldest cores of continents (Chapman and Pollock, 1977; Artemieva and Mooney, 2001), has the longest subsidence histories.

Most models of thermal reequilibration are similar in approach to that used by Haxby et al. (1976) for the Michigan Basin. In this model a large-scale, but undocumented, thermal event leads to formation of a dense crustal mass that causes subsequent subsidence as the lithosphere cools. Flexure broadens the width of basin deflection (Nunn and Sleep, 1984).

Notable in the subsidence curves are the deviations from idealized thermal subsidence (Fig. 5). These deviations are more pronounced than those seen in passive margins and suggest that tectonic reactivation characterizes many intracontinental basins. The most extreme of these is the Ordos Basin (Fig. 4, line 6). However, recent work suggests that this basin may be the result of constructive interaction of deformation events around the basin margin and is not primarily driven by thermal effects (Xie, 2007). Most of the other examples also show strong deviations away from simple thermal equilibration, more than can be reasonably accounted for by eustatic sea-level changes. Various authors have suggested interacting tectonic mechanisms impacting these basins including intraplate stresses, multiple thermal perturbations, reactivation of inherited structures, far-field effects of nearby tectonic events, or changes in lithosphere rheology (Nunn and Sleep, 1984; Klein and Hsui, 1987; Bond, 1991; Kaminski and Jau-part, 2000).

Foreland Basins (Fig. 6)

Foreland basins are asymmetric basins adjacent, and parallel, to an attendant contractional orogenic belt. Foreland basins, or foredeeps, sit atop a deflected continental lithosphere of the underlying plate in both continental collision zones (peripheral foreland basins of Dickinson [1976]) and behind volcanic arcs (retroarc foreland basin of Dickinson [1976]). Many studies have demonstrated that, for the most part, these basins form as a regional isostatic (flexural) response to loading by the adjacent orogenic belt (e.g., Beaumont, 1981; Jordan, 1982; DeCelles and Giles, 1996).

Foreland basin subsidence curves differ from thermal subsidence curves seen in most other basins in that the former are characterized by their convex-up shape and frequent episodic subsidence events. The convex-up profile reflects accelerating subsidence as the tectonic load migrates toward the foreland coupled with the curved flexural profile of the basin. As the basin widens due to migration of the thrust load and associated sedimentation, the distal parts of the

STRIKE-SLIP BASINS

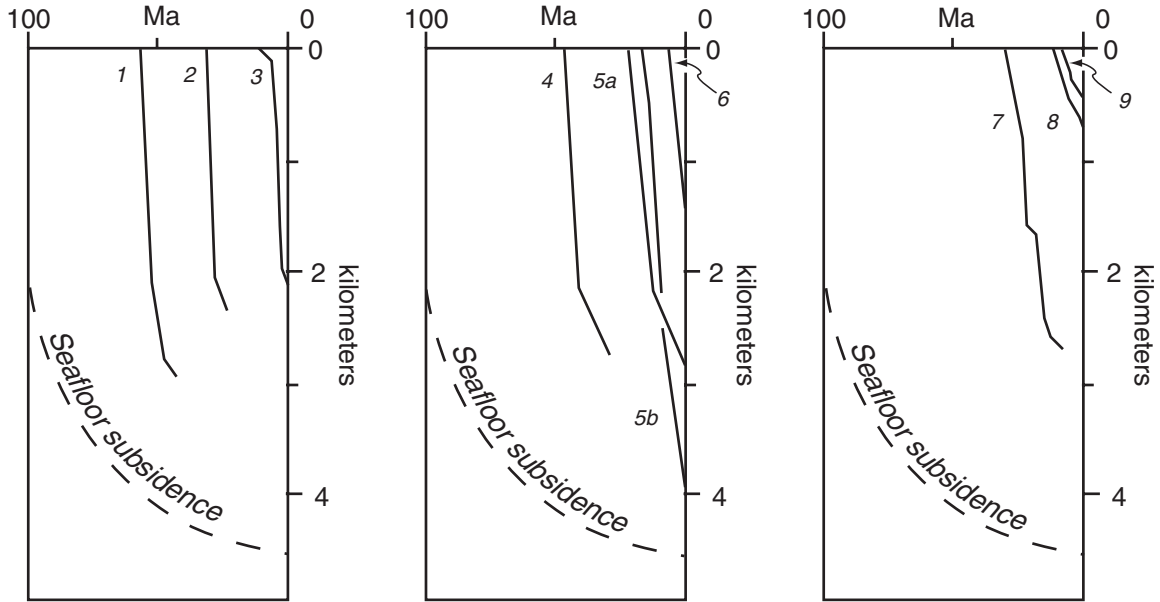


Figure 3. Tectonic subsidence curves for strike-slip basins. Locations shown in Figure 1. Thermal decay curve (dashed) for subsidence of cooling seafloor (Stein and Stein, 1992), minus 500 m, is shown for comparison. 1—Chuckanut Basin (Johnson, 1984, 1985); 2—Ridge Basin (Crowell and Link, 1982; Karner and Dewey, 1986); 3—Death Valley (Hunt and Mabey, 1966); 4—Salinian block (Graham, 1976); 5—Los Angeles Basin (Rumelhart and Ingersoll, 1997); 6—Gulf of California (Curry and Moore, 1984); 7—Cuyama Basin (Dickinson et al., 1987); 8—Bozhang Depression (Hu et al., 2001); 9—Salton Trough (Kerr et al., 1979).

INTRACONTINENTAL BASINS

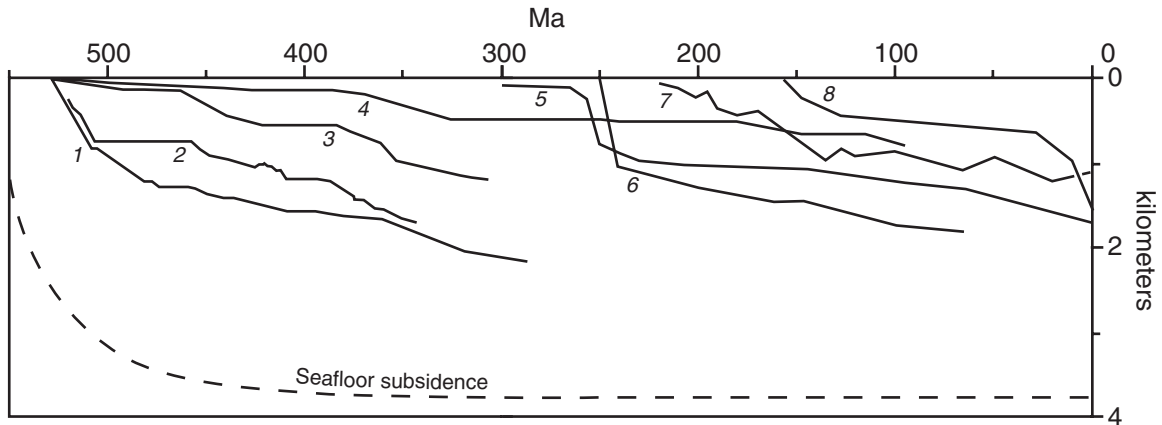


Figure 4. Tectonic subsidence of intracontinental basins. Locations shown in Figure 1. See thermal decay curve (dashed) for subsidence of cooling seafloor (Stein and Stein, 1992), minus 1500 m, is shown for comparison. 1—Illinois Basin, Farley well (Bond and Kominz, 1984); 2—Michigan Basin (Bond and Kominz, 1984); 3—Williston Basin, North Dakota (Bond and Kominz, 1984); 4—Williston Basin, Saskatchewan (Fowler and Nisbet, 1985); 5—Northeast German Basin (Scheck and Bayer, 1999); 6—Southwest Ordos Basin (Xie, 2007); 7—Paris Basin (Prijs et al., 2000); 8—Parana Basin (Zalan et al., 1990).

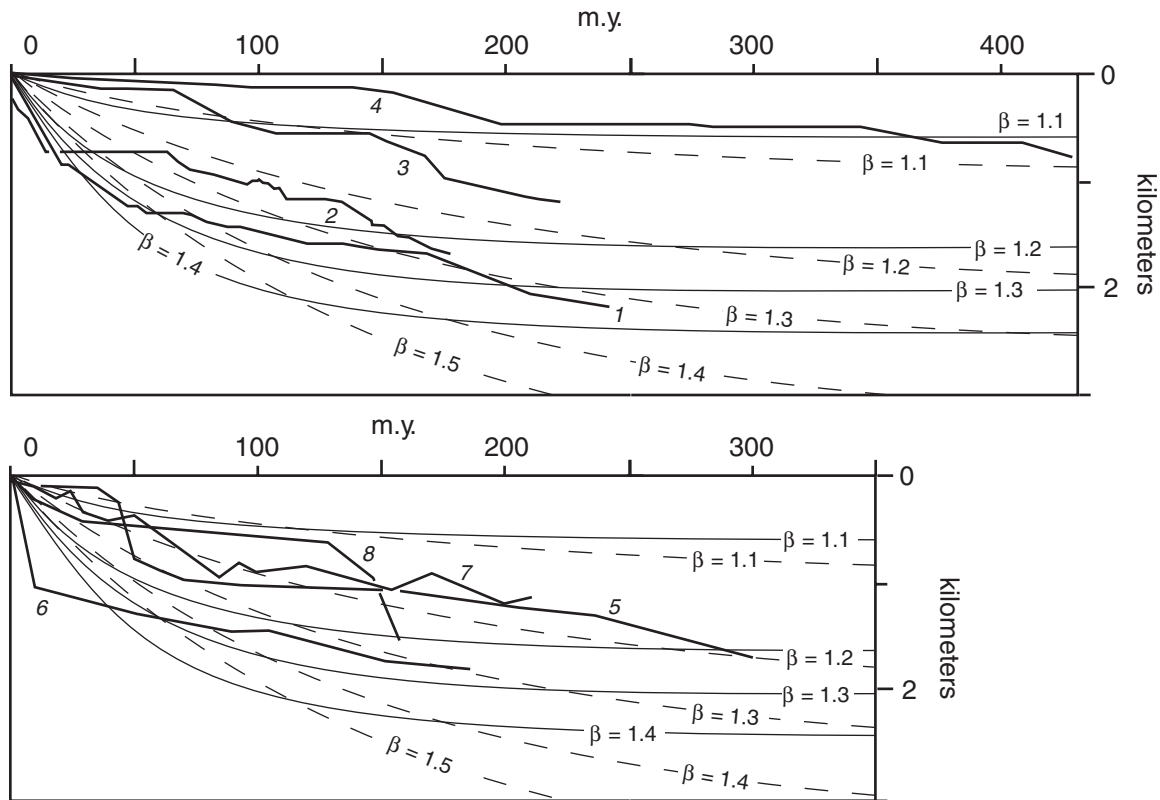


Figure 5. Comparison of intracontinental basin subsidence curves (numbered heavy lines from Fig. 4) with post-rift thermal subsidence curves calculated from the McKenzie (1978) stretching model. Time (m.y.) is shown since basin formation. Thin solid lines assume lithosphere thickness of 125 km; dashed lines assume lithosphere thickness of 200 km. Stretching factors (β) from 1.1 to 1.5 are shown.

basin show time-transgressive subsidence. That is, while the proximal foreland basin responds immediately to adjacent thrust loads, the distal parts of the basin may show later subsidence as loads migrate basinward over time (Jones et al., 2004). The result is a time lag as the tectonic and sediment load propagates across a foreland basin. In addition, the redistribution of sediment and subsidence over time in this way leads to flattening of the basin geometry and reduction in size of the attendant forebulge. Jones et al. (2004) suggest time lags of tectonic signals on the order of a few million years or less out across foreland basins. The total duration of orogens, as seen in subsidence histories (Fig. 6), is typically a few tens of millions of years.

Smaller scale episodes of subsidence superimposed on the overall subsidence profile primarily reflect intermittent thrust events (e.g., Heller et al., 1986), although not every event significantly changes the configuration of the thrust load. Duration and episodicity of subsidence varies from basin to basin, as set by the pace of growth of the adjacent orogen, and in

different parts of a single basin (e.g., Fig. 6, lines 8a and 8b), as a function of local loading history. The maximum magnitude of tectonic subsidence seen in compiled curves is ~ 3 km.

Blind thrusts often propagate into the proximal foreland basin and sedimentation can continue above these structures. DeCelles and Giles (1996) refer to this part of the foreland basin as the “wedge top.” While these basins are not included here, it is clear that thrust emplacement will impact subsidence history in these parts of the proximal foreland (e.g., Vergés et al., 1998). Other smaller basins may form in concert with deformation of the adjacent orogen. These include piggyback and back-bulge basins (Ori and Friend, 1984; DeCelles and Giles, 1996). Subsidence history of piggyback basins tends to be shorter lived and of less magnitude than their associated foreland basins (e.g., Burbank et al., 1992; Carrapa et al., 2003), and is not considered here. Back-bulge basins, if present, are very subdued features that lie outboard of foreland basins, beyond the forebulge, and form as a dampened flexural response to loading

of an elastic plate. Magnitude of deflection of back-bulge basins is very small, typically a few percent of the depth of the associated foreland basin, and may be difficult to uniquely identify.

Forearc Basins (Fig. 7)

Forearc basins lie between trenches and their associated, parallel, magmatic arcs (Dickinson, 1995). The sizes and configurations of both modern and ancient forearc basins are highly variable, but it is clear that typical forearc basins are narrow and elongate, with thick sediment packages confined to deep structural troughs. We note that there is often a large range of paleobathymetry in these settings, so that resultant subsidence curves may be less well constrained.

Subsidence curves from forearc basins, as a group, have the most diverse range of shapes (Fig. 7). Some show very rapid, short-lived subsidence similar to strike-slip basins. Others have slower, relatively linear subsidence. Still others show an abrupt transition from rapid subsidence to very slow subsidence rates, similar to some

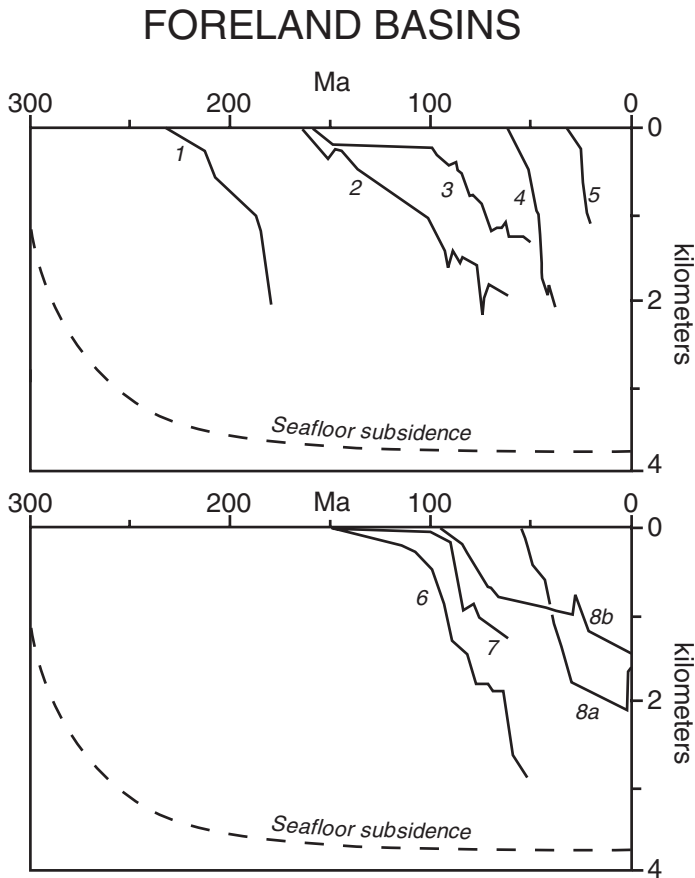


Figure 6. Tectonic subsidence of foreland basins. Locations shown in Figure 1. Thermal decay curve (dashed) for subsidence of cooling seafloor (Stein and Stein, 1992), minus 1500 m, is shown for comparison. 1—Eastern Avalonia, Anglo-Brabant fold belts (van Grootel et al., 1997); 2—Southern Alberta Basin (Gillespie and Heller, 1995); 3—San Rafael Swell, Utah (Heller et al., 1986); 4—Pyrenean foreland basin, Gombren (Vergés et al., 1998); 5—Swiss Molasse basin (Burkhard and Sommaruga, 1998) modified from total subsidence using water:sediment density contrast); 6—Hoback Basin, Wyoming (Cross, 1986); 7—Green River Basin, Wyoming (Cross, 1986; Heller et al., 1986); 8—Magallanes Basin (Biddle et al., 1986).

the underlying subducted slab that isostatically impact the overlying plate (Kobayashi, 1995), and collision of crustal fragments in the subduction zone (Clift and MacLeod, 1999).

Possible subsidence mechanisms in forearc basins include growth, loading, and underplating of the accretionary prism, which may drive tectonic rotation and basin widening in some settings (Coulbourn and Moberly, 1977). Basin growth has also been tied to an increase in width of the arc-trench gap due to flattening of the underthrust plate and resultant migration of the accretionary wedge and volcanic arc (Dickinson, 1995). Regional isostatic effects of changing lithospheric thickness and density due to age and structure of the underthrust plate can account for segmentation, and even uplift, of forearc basin subsidence (Moxon and Graham, 1987). Of course, compression associated with the coupling of the upper and lower plates across convergent margins suggests that folding and thrust loading may contribute to subsidence (Fuller et al., 2006). However, extensional faulting may contribute to subsidence in some forearc settings (Izart et al., 1994; Unruh et al., 2007). Thermal subsidence associated with either cooling of the flank of the adjacent arc massif (Moxon and Graham, 1987) or cooling of an accreted warm microplate (Angevine et al., 1990) are possible mechanisms. In fact, thermal subsidence in forearc settings can be accelerated due to refrigeration by the underthrust plate (Mikhailov et al., 2007). Clift and MacLeod (1999) discuss the role of tectonic erosion by the down-going slab as a cause of subsidence and tilting of the forearc basin. Subsidence of forearc basins is the least understood and most poorly constrained of the tectonic settings explored in this study.

SUMMARY

Tectonic setting exerts primary control on sedimentary basin subsidence history. Several basin settings seem to have distinctive subsidence patterns suggesting a limited range of driving mechanisms. As such, calculated subsidence history is a potential tool for identifying tectonic setting of ancient basins of unknown origin. Passive margins show rapid initial synrift subsidence followed by prolonged thermal subsidence similar to that seen for subsiding seafloor. Strike-slip basins all demonstrate rapid, albeit short-lived, subsidence. Foreland basins are characterized by segmented convex-up subsidence. Intracontinental basins studied here show long-lived gradual subsidence. While overall the subsidence pattern of intracontinental basins is consistent with thermal subsidence of thick lithosphere, most profiles contain large deviations from predicted

curves from passive margins. In addition, some forearc basins show large uplift events, such as in the Indonesian forearc basin (Beaudry and Moore, 1985). Other basins, such as the Chilean forearc, show significant amounts of rotation and widening of basin fills over time (Coulbourn and Moberly, 1977). Most of the curves show less than 2 km of tectonic subsidence. The modern Tonga forearc is exceptionally deep (Fig. 7, line 4).

The range of shapes of subsidence curves in this setting indicates that a variety of factors may contribute to forearc basin subsidence. Most curves are relatively simple in form and imply a monotonic driving mechanism. The curves

from the Great Valley of California (Fig. 7, line 1), exhibit an abrupt change in subsidence rate possibly reflective of a change in driving mechanism. The Great Valley curves also show very different timings of inflection points indicating that the basin is tectonically segmented into differentially subsiding zones. Basin segmentation is seen elsewhere (Izart et al., 1994) and may be common in these settings. Episodic subsidence and even uplift of some basins is seen in Figure 7, although it is not clear to what extent these may reflect errors in bathymetric assignments. Causes of segmentation include partitioned strain associated with oblique subduction (Izart et al., 1994), bathymetric changes in

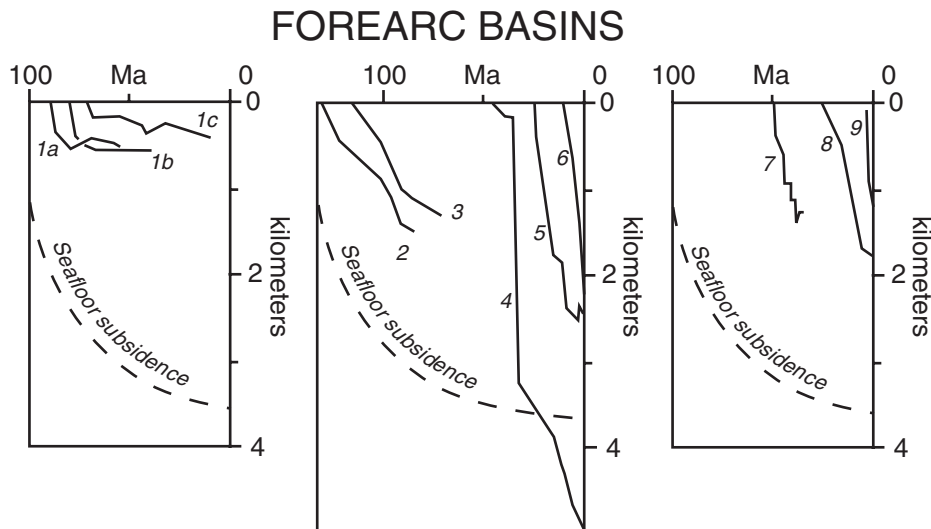


Figure 7. Tectonic subsidence curves of forearc basins. Locations shown in Figure 1. Thermal decay curve (dashed) for subsidence of cooling seafloor (Stein and Stein, 1992), minus 1500 m, is shown for comparison. 1—Great Valley (Moxon and Graham, 1987); 2—Sacramento Basin (Dickinson et al., 1987); 3—Peninsular Ranges (Kimbrough et al., 2001); 4—Tonga forearc (Clift and MacLeod, 1999); 5—Japan forearc, Deep Sea Drilling Project sites 438 and 439 (von Huene, 1982; Ingle, 1992); 6—Southern Lesser Antilles, northwest of Tabago (Ysaccis, 1997); 7—Oregon Coast Range (Angevine et al., 1990; Heller, 1983); 8—Luzon Central Valley (Bachman et al., 1983); 9—Kazusa forearc basin (Ito, 1995).

thermal curves suggestive of tectonic reactivation. Forearc basins studied have various subsidence profiles, suggesting there may be a range of driving mechanisms of subsidence.

Although subsidence analysis can be a useful tool in identifying tectonic setting in ancient sequences, second-order variations in the subsidence rate provides specific information on important local details of driving forces and tectonic timing. Finally, caution should be used given the limitation of data sets available for this compilation and because basins can span multiple tectonic settings over time and/or space. This approach is best used in conjunction with other structural and basin analysis techniques.

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