Tectonic subsidence of the early Paleozoic passive continental margin in eastern California and southern Nevada

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ABSTRACT

Quantitative analysis of tectonic subsidence in Cambrian and Ordovician platform carbonates and associated strata exposed in the Spring Mountains (Nevada) and the Nopah, Funeral, and Inyo Ranges (California) indicates that subsidence associated with this segment of the early Paleozoic passive continental margin is exponential in form, consistent with thermal contraction of the lithosphere following extension. As in other parts of the North American Cordillera, continental separation in the southern Great Basin appears to have taken place between 590 and 545 Ma. These results are not sensitive to uncertainties in stratigraphic thickness, biostratigraphic age control, or paleobathymetry. Uncertainties in the Cambrian time scale lead to predictable variations in the inferred time of onset of thermal subsidence, but they have no effect on the inferred stratigraphic position of the rift to post-rift transition. A younger age for the base of the Middle Cambrian results in a younger inferred age of onset of thermal subsidence accompanied by greater rates of subsidence during the Cambrian, whereas a significantly older estimate of the onset of thermal subsidence can be obtained only if the base of the Middle Cambrian is substantially older than 540 Ma, a possibility that is inconsistent with available data.

Results of the subsidence analysis are particularly significant because this is one of the few regions along the length of the North American Cordillera where they can be compared directly to the geologic evidence for syn-rift and post-rift deposition. Basement-involved faulting associated with the Amargosa basin ("aulacogen") ceased during deposition of the Noonday Dolomite, which is thought to be older than 700–680 Ma on the basis of stromatolites of late Riphean affinity. The overlying Johnnie Formation contains supposed Vendian stromatolites (younger than 700–680 Ma). If it is assumed that our results indicate the timing of the final rift to post-rift transition, then either the ages inferred from stromatolites are incorrect or the lithosphere was thinned regionally after deposition of the Noonday. The latter possibility is supported by limited geologic evidence for extension in latest Proterozoic and Early Cambrian time. The lack of appreciable physical evidence for crustal extension after deposition of the Noonday, however, may imply that (1) a uniform extension model for lithospheric thinning is inappropriate for this part of the margin or that (2) some or all of the localities studied are continentward of the hinge zone, and that the observed subsidence is exaggerated by flexural loading in a deeper basin to the west.

INTRODUCTION

Evidence for fragmentation of the Laurentian supercontinent during Late Proterozoic and early Paleozoic time is widespread. Grabens and passive continental margins of this age are preserved on many continents (Bond and others, 1984), and continental dispersal appears to have been accompanied by a global rise of sea level related to a decrease in the mean age of the oceanic crust (Bond and others, 1988, 1989). Critical to attempts to reconstruct the Laurentian supercontinent is determining precisely the time of breakup for specific passive-margin segments. Counterpart margins may or may not have a common history prior to breakup, depending on the distribution of older sutures for example, but they should be of the same age. Following earlier work in Mesozoic and Cenozoic margins (for example, Watts and Ryan, 1976; Steckler and Watts, 1978, 1982), Bond and Kominz (1984) developed a procedure to recover tectonic subsidence from fully lithified rocks. Tectonic subsidence is the subsidence that would occur in a sedimentary basin in the absence of sedimentation and eustatic variations. Quantitative studies of tectonic subsidence associated with formation of the early Paleozoic passive margins for the North American Cordillera (Bond and others, 1983, 1984, 1985; Armin and Mayer, 1983; Bond and Kominz, 1984), eastern North America (Bond and others, 1984), several basins across Australia (Bond and others, 1984; Lindsay and others, 1987), southeast Turkey, and northwest Argentina (Bond and others, 1984) suggest that in many places rifting ceased and passive margins formed during latest Proterozoic and Cambrian time. The age of onset of thermal subsidence indicated by these studies ranges from about 600 to 555 Ma.

In contrast to the timing of passive-margin development indicated by quantitative subsidence analysis, the most convincing geologic evidence for the main rifting event in the western United States is present in strata between ~800 and 700 Ma (Stewart, 1972; Wright and others, 1976; Stewart and Suczek, 1977; Link, 1984; Miller, 1985, 1987; Link and others, 1987). The precise stratigraphic position of the transition from rift-related to passive-margin sedimentation remains controversial, however, with the range of estimates spanning an interval more than 2 km thick (Stewart, 1972, 1976; Stewart and Poole, 1974; Wright and others, 1976; Stewart and Suczek, 1977; Christie-Blick, 1984; Link, 1984; Bond and others, 1985; Prave and Wright, 1986; Link and others, 1987; Christie-Blick and Levy, 1989a).

The purpose of this paper is to apply quantitative subsidence analysis to the early Paleozoic passive continental margin of the southern Great

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Basin (Fig. 1), updating earlier studies by Stewart and Suczek (1977) and Armin and Mayer (1983), and to address the apparent discrepancy between the geologic evidence for the timing of rifting and the results of subsidence analysis, as the subsidence has been analyzed for the most part from segments of the margin to the north. The area of eastern California and southern Nevada chosen for this study is one of the few segments of the North American Cordillera with appropriate stratigraphy and for which tectonic subsidence has not been studied in detail. Direct evidence for rifting is also well displayed, and tuffaceous beds at several horizons may be amenable to U-Pb geochronology of zircons (A. P. LeHuray and S. A. Bowring, unpub. data).

REGIONAL STRATIGRAPHY AND TECTONIC SETTING

It is generally recognized that strata of Late Proterozoic and early Paleozoic age in the western United States record a transition from intracontinental rifting to the development of a passive continental margin (Stewart, 1972, 1976; Burchfiel and Davis, 1975; Dickinson, 1977; Bond and others, 1985). Marine and nonmarine predominantly siliciclastic sedimentary and volcanic rocks of Late Proterozoic and Early Cambrian age form the lower part of a miogeoclinal wedge and generally thicken westward from <150 m along the eastern margin of the Great Basin to >6,000 m over a distance of ~200 to 300 km (Misch and Hazzard, 1962; Stewart and Poole, 1974). The upper part of this wedge consists for the most part of peritidal and subtidal carbonate rocks and mudstones of Middle Cambrian to Devonian age and is as much as 5,000 m thick (Stewart and Poole, 1974).

Strata of Middle to Late Proterozoic Age

Stratigraphy. The oldest sedimentary rocks in eastern California and southern Nevada belong to the Pahrump Group and consist of more than 3,000 m of predominantly fluvial to marine siliciclastic rocks and peritidal carbonate rocks (Fig. 2; Wright and others, 1976, 1981; Labotka and Albee, 1977). The Pahrump Group is divided into three formations: the Crystal Spring Formation at the base (Roberts, 1976, 1982; Maud, 1983), the Beck Spring Dolomite (Gutstadt, 1968; Marian, 1979; Tucker, 1983; Zempolich and others, 1988), and the Kingston Peak Formation at the top (J.M.G. Miller and others, 1981; Troxel, 1982; Miller, 1985; Walker and others, 1986). The Crystal Spring Formation unconformably overlies crystalline basement with U-Pb ages of 1.8 to 1.4 Ga (Wasserburg and others, 1959; Silver and others, 1961; Lanphere and others, 1964; Stern and others, 1966; Labotka and Albee, 1977; Labotka and others, 1980; Dewitt and others, 1984) and contains *Baicalia*-type stromatolites of middle Riphean affinity, suggesting an age of 1.35 to 0.95 Ga (Cloud and Semikhatov, 1969; Raaben, 1969; Roberts, 1982). Diabase sills in the Crystal Spring Formation have been tentatively correlated with 1.1–1.2 Ga sills in Arizona (Wrucke and Shride, 1972; Spall and Troxel, 1974; Roberts, 1982). The Kingston Peak Formation contains diamictite, in part of glacial origin. These strata have been correlated with similar rocks at numerous localities in the Cordillera (Crittenden and others, 1972; Christie-Blick and others, 1980) and are best dated as between 770 and 720 Ma (Armstrong and others, 1982; Evenchick and others, 1984; Devlin and others, 1985, 1988; Roots and Parrish, 1988).

The Pahrump Group is conformably to unconformably overlain by

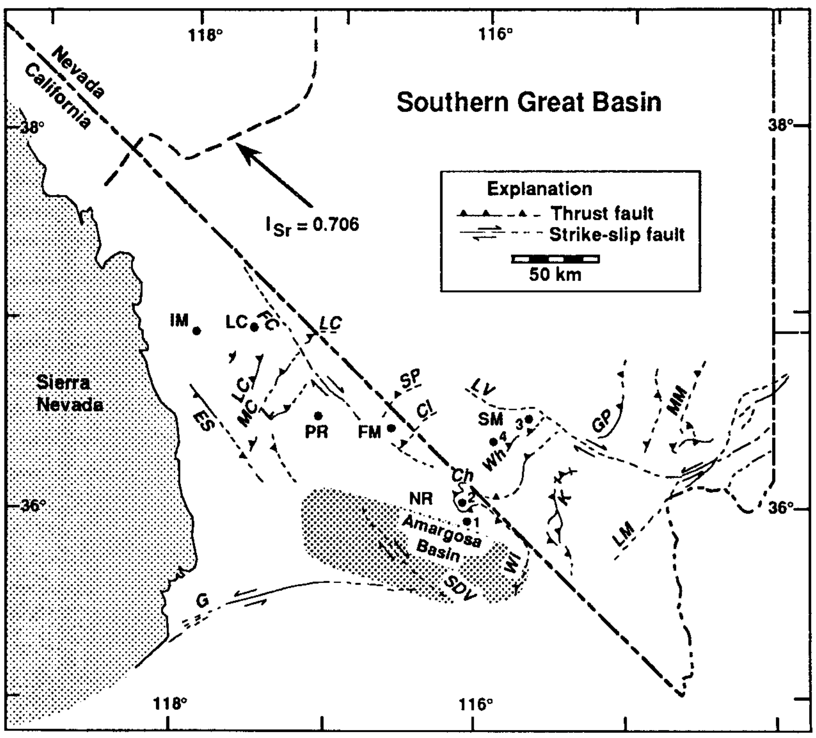

Figure 1. Map of eastern California and southern Nevada, showing locations of stratigraphic sections (dots; modified from Levy and Christie-Blick, 1989): FM, Funeral Mountains (Pyramid Peak, Echo Canyon); IM, Inyo Mountains (Mazourka Canyon); LC, Last Chance Range; NR, Nopah Range (1, Emigrant Pass—Carrara Formation; 2, west side of Nopah Range—Bonanza King Formation through Ely Springs Dolomite); PR, Panamint Range; SM, Spring Mountains (3, Indian Ridge—Bonanza King and Nopah Formations; 4, Wheeler Wash—Carrara Formation and Pogonip Group through Ely Springs Dolomite). On the basis of a recent interpretation of the structure of the Nopah Range (Wernicke and others, 1988a, 1988b), the two sections in the Nopah Range are from the Chicago Pass and Keystone thrust plates, respectively. After palinspastic reconstruction, these two sections are approximately 12 km farther apart (see Fig. 3; Levy and Christie-Blick, 1989a). Faults: *Cl*, Clery thrust; *Ch*, Chicago Pass thrust; *ES*, East Sierran thrust system; *FC*, Northern Death Valley–Furnace Creek fault zone; *G*, Garlock fault; *GP*, Gass Peak thrust; *K*, Keystone thrust; *L*, Lemoigne thrust; *LC*, Last Chance thrust; *LM*, Lake Meade fault system; *LV*, Las Vegas Valley shear zone; *MC*, Marble Canyon; *MM*, Muddy Mountains thrust; *SDV*, Southern Death Valley fault; *SP*, Schwaub Peak thrust; *Wh*, Wheeler Pass thrust; *Wi*, Winters Pass thrust. The dashed I_{Sr} = 0.706 line is from Kistler and Lee (1989).

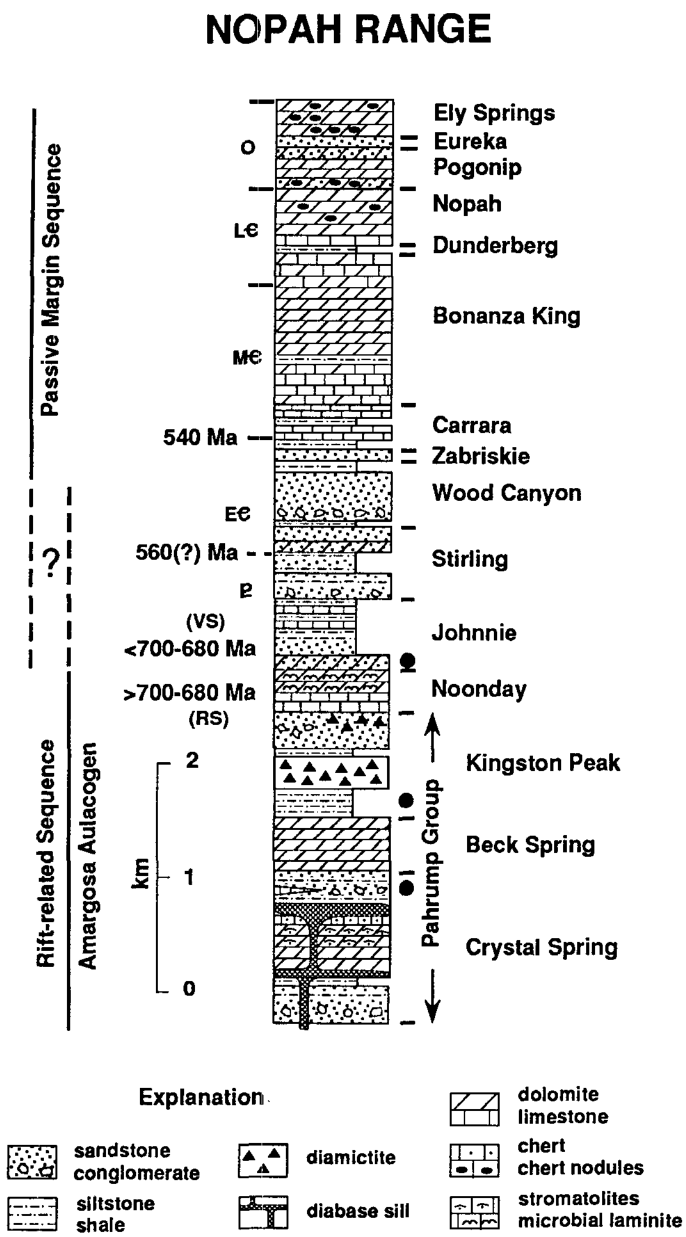

Figure 2. Simplified stratigraphic column for the Nopah Range and adjacent areas from the Middle Proterozoic through the Ordovician (Hazzard, 1937; Wright, 1973; Wright and others, 1976; Halley, 1974; Miller, 1982; Palmer and Halley, 1979; Burchfiel and others, 1982; Sundberg, 1982). The Crystal Spring Formation through Zabriske Quartzite are generalized from the Death Valley region (Palmer, 1971; Wright and others, 1976; Prave, 1984; Wertz, 1982, 1984). The stratigraphic position of the rift to post-rift transition is thought to be located within the uppermost Proterozoic to Lower Cambrian rocks, delineated by the dashed lines to the left of the column. VS, Vendian stromatolites; RS, Riphean stromatolites; black dots, tuffaceous horizons. Base of Middle Cambrian, 540 Ma (Harland and others, 1982); base of Cambrian, 560(?) Ma on the basis of the age of the Mistaken Point Formation (Benus, 1988); stromatolite ages from Cloud and Semikhatov (1969) and Raaben (1969).

the Noonday Dolomite, which consists of as much as 400 m of stromatolitic dolomite and changes facies abruptly toward the south into a deeper-water basinal equivalent (Cloud and others, 1974; Wright and others, 1976, 1978; Williams and others, 1976; Wright and Troxel, 1984). Stromatolites in the Noonday Dolomite are considered by some to be similar to upper Riphean stromatolites of Siberia, suggesting an age of >700–680 Ma (Wright and others, 1978; Miller, 1985, 1987).

Tectonic Setting. Wright and others (1976) proposed that the Pahrump Group was deposited in a fault-bounded basin, the Amargosa "aulacogen." Evidence for the existence of this fault-bounded basin includes facies, thickness, and clast size trends; compositional data and paleocurrent trends, which appear to indicate derivation from elevated crustal blocks to the north and south; and the presence of mafic and felsic volcanic rocks in both the Crystal Spring and Kingston Peak Formations (Roberts, 1976, 1982; Wright and others, 1976; Stewart and Suczek, 1977; Maud, 1983; Miller, 1985; Walker and others, 1986, 1987; Troxel and others, 1987). The strongest evidence for rifting is present in the Kingston Peak Formation and lower part of the overlying Noonday Dolomite (Wright and others, 1976; Stewart and Suczek, 1977; Miller, 1985; Walker and others, 1986), but stratigraphic evidence also indicates at least two distinct intervals of crustal extension during accumulation of the Crystal Spring Formation (Roberts, 1976, 1982).

The Noonday Dolomite, deposited on a fault-bounded platform along the northern margin of the Amargosa basin, unconformably overlies the erosional edges of each of the formations of the Pahrump Group and in places rests directly on the crystalline basement (Wright and others, 1976, 1978; Williams and others, 1976). It is of tectonic significance because it appears to span a transition from fault-controlled subsidence to a phase of thermally driven subsidence (Miller, 1987), but appears to predate by at least 100 m.y. the onset of the main phase of thermal subsidence associated with the Paleozoic passive continental margin.

Although stratigraphic evidence for the existence of the Amargosa basin is compelling (Wright and others, 1976), new evidence is now emerging that the apparent eastern trend of this basin may be largely a function of pronounced extension in this direction during late Cenozoic time (Wernicke and others, 1988a, 1988b). Restoration of this deformation dramatically foreshortens the basin (Levy and Christie-Blick, 1989) and opens the possibility that its original orientation was quite different, perhaps consistent with the north-south stratigraphic trends seen in the Panamint Range (PR in Fig. 1; Labotka and Albee, 1977; Miller, 1985).

Strata of Latest Proterozoic to Early Cambrian Age

Stratigraphy. Overlying the Noonday Dolomite is a relatively conformable succession, as much as several kilometers thick, of predominantly siliciclastic shallow-marine to fluvial rocks, with minor amounts of carbonate and mafic volcanic rocks (Fig. 2; Hazzard, 1937; Stewart, 1970, 1974; Wright and others, 1981). Lithological and paleoecological correlation with sections in the White-Inyo region (IM in Fig. 1) suggests that the base of the Cambrian section lies within the Stirling Quartzite, possibly near the base of the dolomitic D member (Fig. 2; Stewart, 1970, 1982; Nelson, 1976; Signor and Mount, 1989), a unit in which problematic shelly fossils presumably of Cambrian age have been found (Langille, 1974). In general, however, the age of these rocks is poorly constrained.

Tectonic Setting. It has been suggested that the transition from intracontinental rifting to continental separation and passive-margin formation occurred during deposition of these strata (Bond and others, 1983, 1985; Armin and Mayer, 1983; Bond and Kominz, 1984; Christie-Blick,

TABLE 1: BIOSTRATIGRAPHIC AND THICKNESS DATA

Global Chronostratigraphic Units			North American Chronostratigraphic Units		Ma	Nopah Range	Spring Mountains	Funeral Mountains	Inyo Mountains
SIL	EARLY	Llandoverian			**428**				
ORDOVICIAN	LATE	Ashgillian	Cincin-Natian	Richmondian	438	Ely Springs Dolomite 329	Ely Springs Dolomite 118	Ely Springs Dolomite 94	Ely Sp Dol: Units 2 & 3 78
				Maysvillian					Ely Sp Dol: Unit 1 81
				Edenian	448				
		Caradocian	Champlainian	Shermanian					Johnson Spring Fm 40
				Kirkfieldian		Eureka Quartzite 81	Eureka Quartzite 77	Eureka Quartzite 171	
				Rocklandian					
				Black-Riverian					
	MIDDLE	Llandeilian		Chazyan	458	Pogonip Group 317	Pogonip Group 904	Pogonip Group 664	Barrel Spring Fm 26
		Llanvirnian		White-Rockian	468				Badger Flat 275
	EARLY	Arenigian	Canadian (Ibexian)		478				Al Rose Fm 135
		Tremadocian		*Missisquoia*	**488**				
CAMBRIAN	LATE		Tremp.	*Saukia*	505	Nopah Fm 515	Nopah Fm 459	Nopah Fm 500	Tamarack Canyon 105
			Franc.	*Idahoia*					Lead Gulch Fm 86
				Taenicephalus					
				Elvinia					
			Dresb.	*Dunderbergia*		Dunderberg Shale 22	Dunderberg Shale 37	Dunderberg Shale 59	
				Prehousia		Bonanza King Fm: Banded Mt Member 300	Bonanza King Fm: Upper Cambrian 300	Bonanza King Fm: Upper Cambrian 214	Bonanza King Fm: Upper Cambrian 200
				Dicanthopyge					
				Aphelaspis					
				Crepicephalus					
				Commonly barren	523				
	MIDDLE		*Eldoradia*			Banded Mt Member 540	Middle Cambrian 1028	Middle Cambrian 700	Middle Cambrian 656
			Commonly barren			Papoose Lake Member 475			
			Ehmaniella						
			Glossopleura			Carrara Fm: Desert Range Ls 39, Jangle Ls 95, Pahrump Hills Sh 50, Red Pass Ls 31, Pyramid Sh 82	Carrara Fm (upper part) 307	Carrara Fm (upper part) 290	Monola Fm 381
			Albertella						
			Plagiura						
	EARLY		*Bonnia-Olenellus*		**540**	Pyramid Sh 10			
			Nevadella						
			Fallotaspis						
			Tommotian		570				

Biostratigraphic and stratigraphic thickness data used in construction of subsidence curves. The geological timescale is from Harland and others (1982). Trilobite zones are from Taylor (1989). Bold numbers on the timescale are tie-points. Bold lines represent boundaries for which biostratigraphic ages can be tied to radiometric ages. Fine solid lines are boundaries for which biostratigraphic ages are not tied to radiometric ages. Dotted lines are boundaries that have uncertain biostratigraphic control or are diachronous. Thickness of units is shown in meters beside formation names. Stratigraphic data used was compiled from our field studies and the following publications. Nopah Range: Hazzard (1937), Wright (1973), Halley (1974), Miller (1982), Palmer and Halley (1979), Burchfiel and others (1982), Sundberg (1982). Spring Mountains: Cornwall (1972), Burchfiel and others (1974). Funeral Mountains: Halley (1974), McAllister (1976). Inyo Mountains: Ross (1964, 1965), Pestana (1960), Miller (1976). Biostratigraphic data are from the following publications: Pestana (1960), Ross (1964, 1965), Palmer (1971), Miller (1976, 1982), Nelson (1976), Sloan (1976), Palmer and Halley (1979), Cooper and others (1982).

1984), but the precise stratigraphic position of the transition is uncertain. Stewart (1974) emphasized the remarkable regional persistence of lithostratigraphic units in this interval, arguing that these rocks therefore represent the earliest deposits of the passive-margin succession, whereas Christie-Blick (1984), Bond and others (1985), and Christie-Blick and Levy (1989a) have suggested that strata within this interval display evidence for syn-depositional extension.

Strata of Middle Cambrian through Ordovician Age

The Middle Cambrian through Ordovician strata consist of more than 3,000 m of platform carbonates with minor amounts of siltstone, sandstone, and quartzite (Table 1; Fig. 2; Hazzard, 1937; Pestana, 1960; Ross, 1964, 1965; Cornwall, 1972; Wright, 1973; Burchfiel and others, 1974; Halley, 1974; McAllister, 1976; Miller, 1976, 1982; Palmer and Halley, 1979; Burchfiel and others, 1982; Sundberg, 1982) and are dated biostratigraphically (Ross, 1965; Palmer, 1971; Miller, 1976, 1982; Palmer and Halley, 1979; Palmer, 1982). The Carrara Formation forms the basal unit of this interval and ranges in age from late Early Cambrian (*Olenellus* zone) to early Middle Cambrian (*Glossopleura* zone; Palmer and Halley, 1979; Palmer, 1982). Within this formation, the base of the Middle Cambrian can be located stratigraphically to within a few tens of meters (Palmer and Halley, 1979; Palmer, 1982). This horizon is crucial to the construction of our subsidence curves because it is the lowest boundary for which it is possible to assign a reliable numerical age. In the Inyo Mountains, the base of the Middle Cambrian coincides approximately with the base of the Monola Formation (Table 1; Palmer, 1971). The Middle Cambrian through Ordovician strata exposed in the Nopah Range, Spring Mountains, and Funeral Mountains were deposited in peritidal environments and pass westward toward the White-Inyo Mountains into deeper-water argillaceous limestones (Kepper, 1981; Cooper and others, 1982; Miller, 1982). Together these rocks are generally regarded as representing deposition on a broad shallow shelf or ramp of a passive continental margin (Halley, 1974; Stewart and Poole, 1974; Kepper, 1976, 1981; Miller, 1976, 1982; Sloan, 1976; Palmer and Halley, 1979; Cooper and others, 1981, 1982; R. H. Miller and others, 1981; Sundberg, 1982).

METHODOLOGY

Tectonic subsidence is calculated for Cambrian and Ordovician strata in the Spring Mountains in Nevada and the Nopah, Funeral, and Inyo Ranges in California (SM, NR, FM, and IM in Fig. 1). These localities were chosen because they contain continuously exposed, relatively undeformed Cambrian and Ordovician strata and are arranged approximately in a transect across the early Paleozoic passive margin (Figs. 1 and 3). No localities were studied between the Funeral Mountains and the Inyo Range because intermediate sections such as the Panamint Range and the Last Chance Range are considered too structurally complex for the type of analysis discussed in this paper (PR and LC in Fig. 1; Hopper, 1947; Albee and others, 1980; Labotka and others, 1980; K. Corbett, 1989, personal commun.). The oldest strata for which tectonic subsidence is analyzed are of Middle Cambrian age. In addition to the lack of reliable age control in rocks older than Middle Cambrian, it is necessary to avoid including syn-rift deposits in the subsidence analysis in order to compare results with models for thermal decay, and strata of Middle Cambrian age and younger are generally regarded as passive-margin deposits. Although older strata are not shown on the subsidence curves, latest Proterozoic and Early Cambrian strata have been delithified because their compaction contributed in part to the subsidence of younger rocks. Strata younger than Ordovician are not included in the study because several unconformities are present in the section, and after about 120 m.y. following the onset of cooling, most of the thermal anomaly should have been dissipated (McKenzie, 1978).

Tectonic subsidence can be recovered from fully lithified lower Paleozoic strata by iteratively correcting measured stratigraphic thicknesses for the effects of compaction, sediment loading, varying paleobathymetry, and eustatic fluctuations to the extent that these are known (Bond and Kominz, 1984). As a first approximation, local isostatic compensation is assumed and the effects of lateral heat flow are ignored because it has been shown that for margins wider than 100 km, ignoring the effects of flexure and lateral heat flow does not alter the form of the subsidence curve, although it does affect the slope of the curve (Beaumont and others, 1982; Steckler and Watts, 1982; Bond and Kominz, 1984; Bond and others, 1988). Palinspastic reconstruction of the southern Great Basin indicates that this part of the margin was at least 200 km wide in early Paleozoic time (Fig. 3; Levy and Christie-Blick, 1989). Therefore, assuming a one-dimensional passive-margin model allows an estimate of the onset of thermal subsidence in the southern Great Basin, although it does not necessarily yield a very good measure of the magnitude of extension (β). Decompaction of strata yields a range of values depending on whether sediments were lithified by physical compaction or by the introduction of externally derived cement. Progressively younger strata are iteratively decompacted and removed from the stratigraphic section (see Fig. 1 of Bond and Kominz, 1984). Decompacted values can vary by no more than 20%–30% between successive data points, and to a first approximation the curve shown is subparallel to the minimum and maximum curves (Fig. 4; Bond and Kominz, 1984; Bond and others, 1988). No correction is made in this study for variations in water depth because facies associations and fossil assemblages indicate that all of the strata accumulated in shallow water. Furthermore, no correction is made for eustatic sea level because at present no reliable interpretation of eustatic change is available for the early Paleozoic. Eliminating this correction results in a subsidence curve that includes exponentially decaying thermal effects, oscillatory eustatic

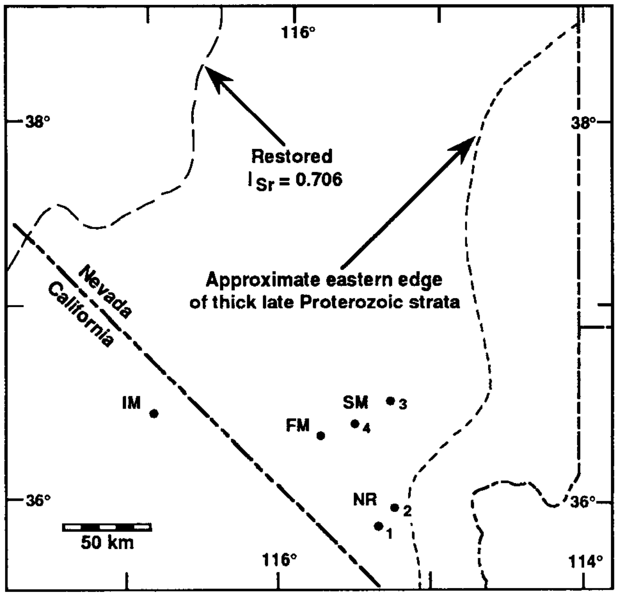

Figure 3. Palinspastic map (modified from Levy and Christie-Blick, 1989) showing the original geographic positions of the stratigraphic sections from the Funeral Mountains (FM), Inyo Mountains (IM), Nopah Range (NR), and Spring Mountains (SM).

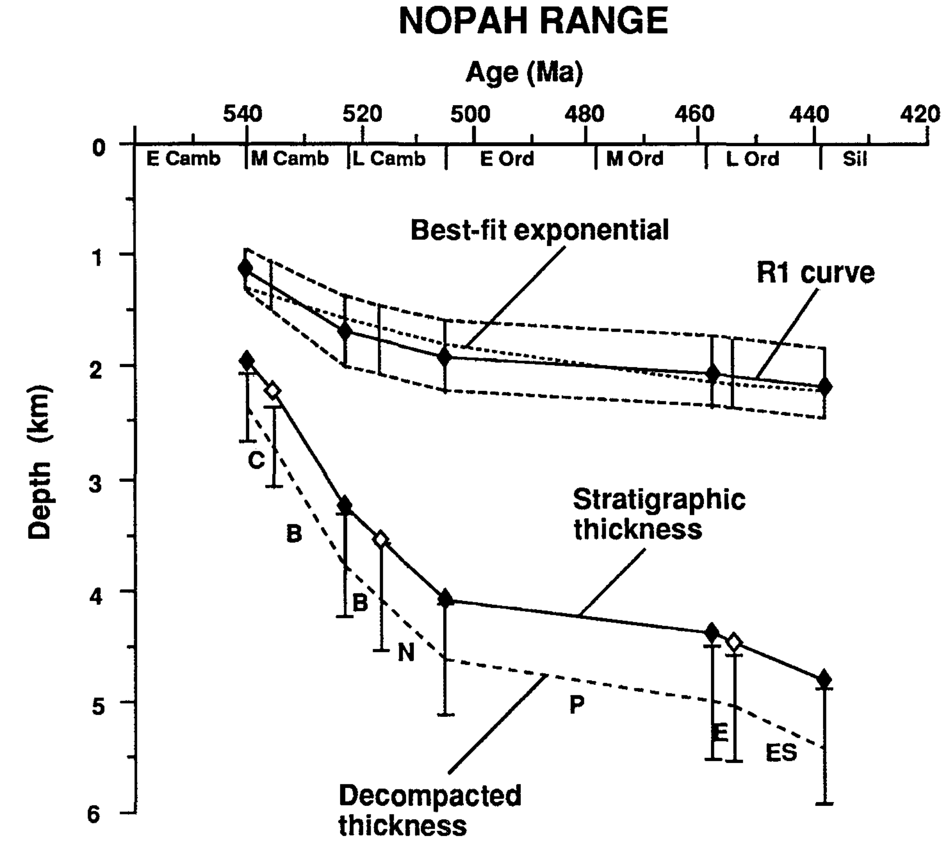

Figure 4. Quantitative analysis of tectonic subsidence of Cambrian and Ordovician rocks of the Nopah Range. The lower solid curve shows present-day stratigraphic thickness plotted as a function of geologic age (time scale of Harland and others, 1982). Strata older than Middle Cambrian are considered poorly dated and are not shown; however, strata of the Johnnie Formation through the lower part of the Carrara Formation (latest Proterozoic through Early Cambrian) have been decompacted, and hence the initial data point includes the thickness of these units. The lower dashed line shows the decompacted thickness of stratigraphic units, with the vertical bars indicating the range of values accounting for lithification as a result of either compaction (maximum value) or cementation (minimum value). The solid diamonds are those formation boundaries that can be tied to radiometric ages; the open diamonds are those horizons for which ages have been interpolated linearly on the basis of present-day stratigraphic thickness. The R1 curve, with delithification ranges, represents the combination of thermal subsidence, long-term eustasy, and randomly varying local effects. The upper dotted line is the best-fit exponential with a decay constant of 62.8 m.y. to the R1 curve and is an approximation of the thermal component of the subsidence. Stratigraphic units: C, Carrara Formation; B, Bonanza King Formation; N, Nopah Formation; P, Pogonip Group; E, Eureka Quartzite; ES, Ely Springs Dolomite.

effects, and randomly varying local effects and that is termed the first reduction of the data or the R1 curve (Fig. 4; Bond and others, 1988; equivalent to Y′ of Watts and Steckler, 1979). In Figure 5, the R1 curves from each of the sections examined for this study are superimposed for reference on the McKenzie (1978) curves.

A best-fit exponential curve with a decay constant of 62.8 m.y., assumed to represent thermal subsidence of ocean floor (McKenzie, 1978), is calculated for the R1 curve by least-squares linear regression and is assumed to approximate the thermal component of the R1 curve (Fig. 4; Bond and others, 1988). The departure of the R1 curve from the best-fit exponential curve indicates eustatic sea-level fluctuations and local effects superimposed on the exponential curve and is used to generate a second reduction of the data, or the R2 curve (Bond and others, 1988). This R2 curve, discussed below, is an approximation of the trend and magnitude of eustatic sea level (Bond and others, 1988).

A linear relation exists between thermally driven subsidence and the square root of time from about 16 to 64 m.y. following rifting that is independent of the model used (Sleep, 1971; McKenzie, 1978; Steckler, 1981), although slight variations may occur when subsidence from earlier rifting (<80 m.y. earlier) enhances the subsidence of the later rifting. Therefore, the R1 subsidence curve and the exponential curve are plotted as a function of the square root of time to determine if the curves are consistent with a thermal mechanism, and to determine the age of onset of thermal subsidence (T_0). In this procedure, a family of R1 curves and best-fit exponentials for a given locality is generated assuming different values of T_0, and the slope of the exponential curves is compared to the rate of ocean-floor subsidence for crust of the same thermal age (Fig. 6; Bond and others, 1983). The maximum age of onset of thermal subsidence coincides with the T_0 of the R1 curve for which the exponential has the greatest slope but does not exceed the rate of ocean-floor subsidence. Regional variations in the rate of subsidence of ocean floor during the first 80 m.y. yield a range of values of $\leqslant$250 to 350 m/m.y.$^{1/2}$ (Parsons and Sclater, 1977; Schroeder, 1984; Hayes, 1988; Marty and Cazenave, 1989). The oldest T_0 estimated for the southern Great Basin, assuming an average rate for ocean-floor subsidence of 300 m/m.y.$^{1/2}$ (Hayes, 1988; Marty and Cazenave, 1989), is <590 Ma (Fig. 6). Inasmuch as thinned continental crust subsides less rapidly than oceanic crust of equivalent thermal age, thermal subsidence may have begun as late as Early Cambrian time (<560 Ma). The younger limit of onset of thermal subsidence is here considered to be 545 Ma on the basis of geologic arguments for the presence of a passive margin by Middle Cambrian time.

ERROR ANALYSIS

A discrepancy exists between the stratigraphic position and age of the strata containing the most convincing geologic evidence for Late Proterozoic rifting (>700 Ma; Wright and others, 1976; Stewart and Suczek, 1977; Miller, 1985; Walker and others, 1986) and the age predicted by the quantitative subsidence analysis for the onset of post-rift thermally driven subsidence (590–545 Ma; Bond and others, 1983, 1985; Armin and

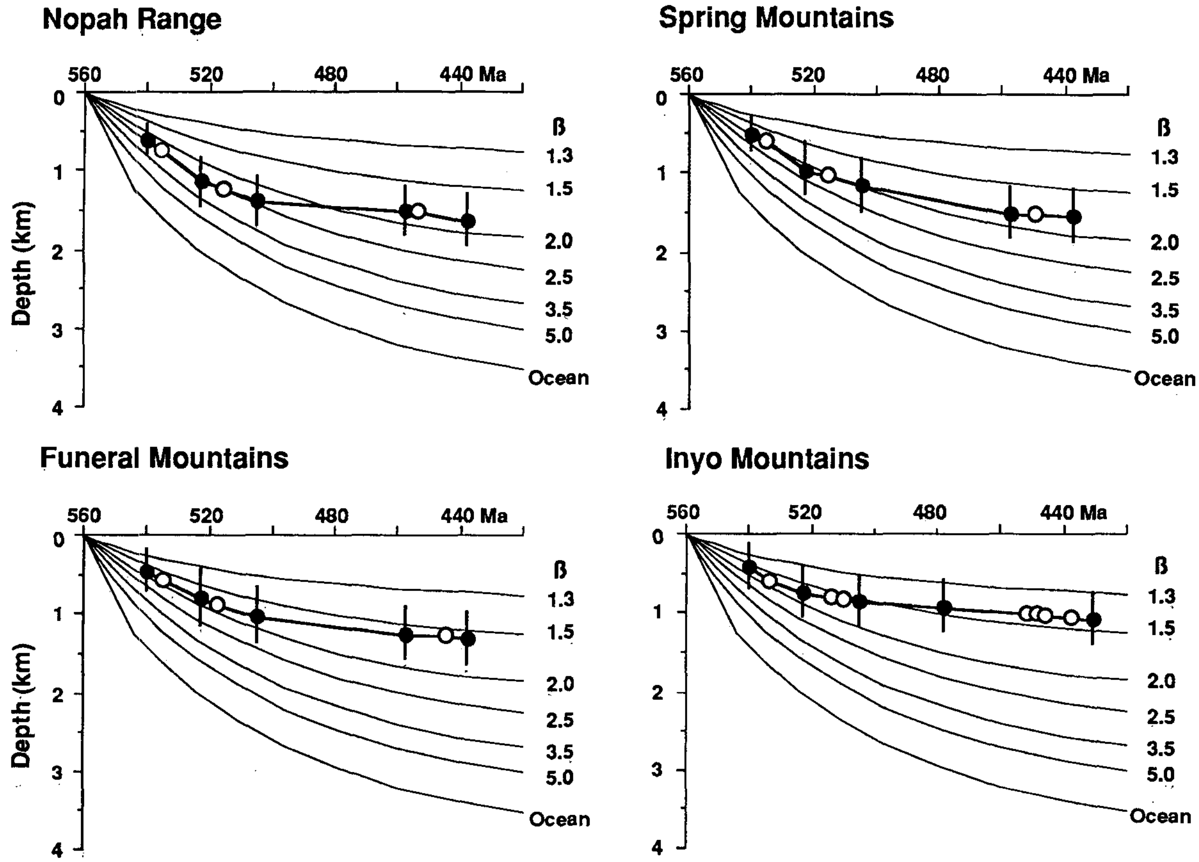

Figure 5. R1 subsidence curves superimposed on the McKenzie (1978) model curves for 125-km lithosphere, assuming an average T_0 of 560 Ma for each locality. Thin curves are the model curves; heavy curves are the R1 curves, with vertical bars representing the range of delithification values; solid circles represent formation boundaries tied to radiometric ages; open circles are located at linearly interpolated formation boundaries; β factors are the apparent stretching values.

Mayer, 1983; Bond and Kominz, 1984; this study). A critical evaluation of the assumptions of the analysis was undertaken to ascertain the sensitivity of results to systematic errors. The Nopah Range was chosen as the reference section because overall it contains the least deformed and most continuous Middle Cambrian through Ordovician succession of the transect. Furthermore, the Nopah Range has been well studied, and therefore abundant published stratigraphic, biostratigraphic, and structural data are available for this locality (Hazzard, 1937; Palmer, 1971; Wright, 1973; Halley, 1974; Miller, 1982; Palmer and Halley, 1979; Burchfiel and others, 1982; Cooper and others, 1982; Sundberg, 1982; McCutcheon and Cooper, 1989; McCutcheon and others, 1989; Griffin, 1989).

Initial Assumptions

Uncertainties in the initial assumptions of the subsidence analysis do not significantly affect the estimate of T_0. These assumptions include (1) local isostatic compensation and one-dimensional heat flow (discussed above), (2) a thermal decay constant of 62.8 m.y. and average ocean-floor subsidence rate of 300 m/m.y.$^{1/2}$, and (3) initial lithospheric thickness of 125 km and crustal thickness of 31 km. The thermal decay constant varies within the modern ocean basins, and it is thus likely that it did so in the Paleozoic oceans. Higher heat flow in the Paleozoic would result in a lower decay constant and faster rates of subsidence. Assuming a decay constant of 40 m.y., lower than observed in modern oceans, reduces the deviation between the R1 curve and the exponential, thus reducing the estimate of the eustatic fluctuations (Bond and others, 1989). Nonetheless, the form of the R2 curve and the timing of the eustatic variations in the early Paleozoic are preserved (Bond and others, 1989). Furthermore, if results of the subsidence analysis are compared with maximum observed ocean-floor subsidence rates in modern oceans of 400 m/m.y.$^{1/2}$ (Marty and Cazenave, 1989), the estimate of T_0 would not be substantially older than 590 Ma (Fig. 6). The rate of subsidence increases to a small extent if the lithospheric thickness is greater, but cannot exceed the rate of ocean-floor subsidence (Bond and others, 1988). The form of the subsidence curve is unaffected by variations in the crustal thickness, although the magnitude of extension necessary to generate a given amount of subsidence is in part dependent on crustal thickness (Bond and others, 1988).

Stratigraphic Thickness

Two potential sources of error in determining accurate stratigraphic thicknesses are related to pressure solution (stylolitization) and undetected faulting. Pressure solution is thought not to have affected the results significantly for two reasons. First, subsidence curves appear to be of thermal form on a regional scale in both eastern and western North America (Bond and others, 1983, 1985; Bond and Kominz, 1984), and their shape is therefore not controlled by random variations in stylolitization. Second, although minor amounts of pressure solution were observed in outcrop

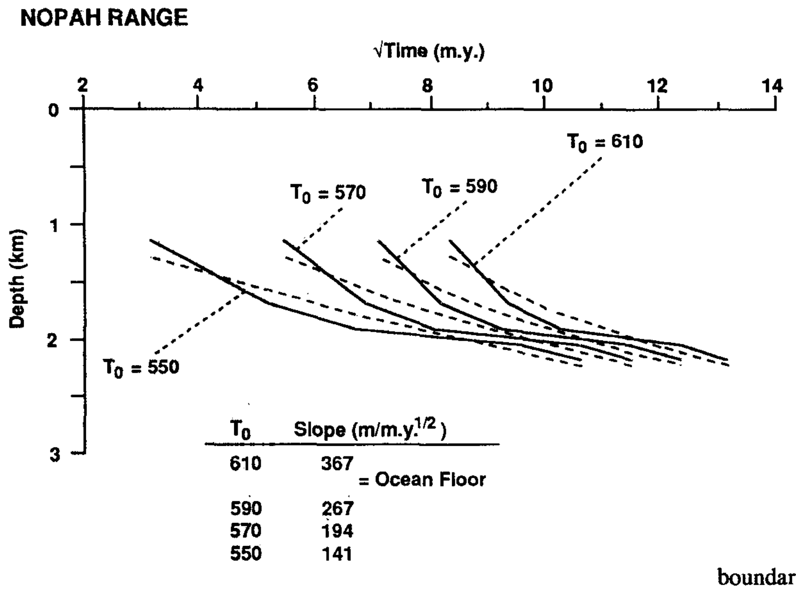

Figure 6. R1 curves and exponential curves plotted as a function of the square root of time for Middle Cambrian through Ordovician rocks of the Nopah Range. A family of curves is generated assuming different values for T_0. Solid lines are the R1 curves; dashed lines are the best-fit exponentials with a decay constant of 62.8 m.y. The slope of the exponential curves is calculated and compared to the rate of ocean-floor subsidence, which ranges from ≤250 to 350 m/m.y.$^{1/2}$ (Parsons and Sclater, 1977; Schroeder, 1984; Hayes, 1988; Marty and Cazenave, 1989). If an average ocean-floor subsidence rate of 300 m/m.y.$^{1/2}$ is assumed, then the oldest estimate of T_0 is <590 Ma. If maximum observed rates of ocean-floor subsidence in modern oceans (400 m/m.y.$^{1/2}$; Marty and Cazenave, 1989) are assumed, then T_0 would be ~610 Ma.

and in thin section, field observation did not indicate a systematic stratigraphic bias in the degree of stylolitization that might have influenced the shape of the subsidence curve. The presence of undetected faults would have the greatest effect on the subsidence analysis if structural complications were preferentially distributed in the Cambrian part of the section where structural thinning would produce a flatter curve and suggest an older estimate of T_0, whereas structural thickening would produce a steeper curve and a younger estimate of T_0.

Through reference to published mapping (Nopah Range—Wright, 1973; Burchfiel and others, 1982, 1983; Spring Mountains—Burchfiel and others, 1974; Funeral Mountains—McAllister, 1971, 1974; Inyo Range—Ross, 1965) and critical re-examination of relevant structural relations in the field, we are confident that only minor structural complications are present. For example, in the Funeral Mountains, pervasive small-scale normal faults were observed in the Bonanza King Formation. On the basis of spacing of and displacement on abundant small-scale normal faults, the apparent map thickness (1,288 m) was estimated to be approximately 30% too great. The corrected stratigraphic thickness (914 m) is shown in Table 1 and Figure 2 and used in the calculation of tectonic subsidence. In the Nopah Range and Spring Mountains, a rather thin section of Carrara Formation is present, possibly as a result of distributed structural thinning. If the true thickness of the Carrara Formation were greater, however, the effect on the subsidence analysis would be to produce a steeper early Middle Cambrian segment of the R1 curve and to indicate a younger inferred timing of onset of thermal subsidence. In the Inyo Mountains, field-checking of structural and stratigraphic data revealed evidence of flattening around Mesozoic plutons, including foliation parallel to the boundaries of the plutons and associated small-scale folds. Although overall flattening appears to be distributed more or less uniformly through the section, the existence of such deformation casts doubt on the significance of the derived R1 curve. The curve is also relatively flat and, for this reason alone, difficult to interpret.

Biostratigraphic Control

Minor uncertainties in the biostratigraphic age of the rocks in this study have little effect on the estimate of T_0. In the Nopah Range, for example, there is little variation in published ages for individual formation boundaries (Palmer, 1971; Palmer and Halley, 1979; Cooper and others, 1982). An exception is the contact between the Nopah Formation and the Pogonip Group, conventionally placed at the Cambrian-Ordovician boundary on the basis of trilobite biostratigraphy (Fig. 2; Hazzard, 1937; Palmer, 1971). Conodont studies indicate that in the Nopah Range, much of the Pogonip Group (defined by Hazzard) is Middle Ordovician (Miller, 1982), although no samples were collected from the lowest 90 m, and an Early Ordovician age for at least part of this interval is not precluded. If the conodont data are accepted, however, linear interpolation of formation boundaries within the Late Cambrian–Early Ordovician interval on the basis of present-day stratigraphic thickness (method of Bond and Kominz, 1984) implies an unrealistically young age (latest Cambrian) for the Dunderberg Shale (Fig. 7B). Available biostratigraphic data place this formation in the *Dunderbergia* biozone (early Late Cambrian; Cooper and others, 1982). This suggests that the Nopah Formation and lowermost Pogonip Group of the Nopah Range accumulated at markedly different rates, possibly owing to the presence of undetected hiatuses, and/or that the Cambrian-Ordovician boundary is located within the interval designated by Hazzard (1937) as Nopah Formation. Consistent with the latter view, Cooper and others (1982) included the upper 235 m of the Nopah Formation (215 m by our measurement) in the Pogonip Group. In our analysis (Table 1; Figs. 4 and 7A), we have followed Burchfiel and others (1982) in accepting Hazzard's lithostratigraphy. None of these uncertainties significantly affects the estimate of T_0 (Fig. 7B).

Biostratigraphic uncertainty also affects the location of the Middle-Late Cambrian boundary within the Bonanza King Formation. At the top of the Papoose Lake Member (lower member of the Bonanza King Formation), the presence of trilobites assigned to *Ehmania*? indicates a Middle Cambrian age (Palmer and Hazzard, 1956; Palmer, 1971). Most of the Banded Mountain Member (upper member) is unfossiliferous, but the upper 100 m has yielded trilobites of the Late Cambrian *Crepicephalus*, *Aphelaspis*, and *Dicanthopyge* zones (Palmer, 1965, 1971). The Middle-Late Cambrian boundary therefore falls somewhere within the unfossiliferous interval. Figures 7C and 7D are subsidence curves generated by assuming a minimum thickness (100 m) and maximum thickness (840 m) of the Banded Mountain Member in the Late Cambrian, respectively (Palmer, 1971). We assume an intermediate thickness of approximately 300 m, or 25%, of the Bonanza King Formation falls within the Late Cambrian (Fig. 7A). The estimate of T_0, however, is not significantly affected by this uncertainty in the stratigraphic position of the Middle-Late Cambrian boundary (Figs. 7A, 7C, and 7D).

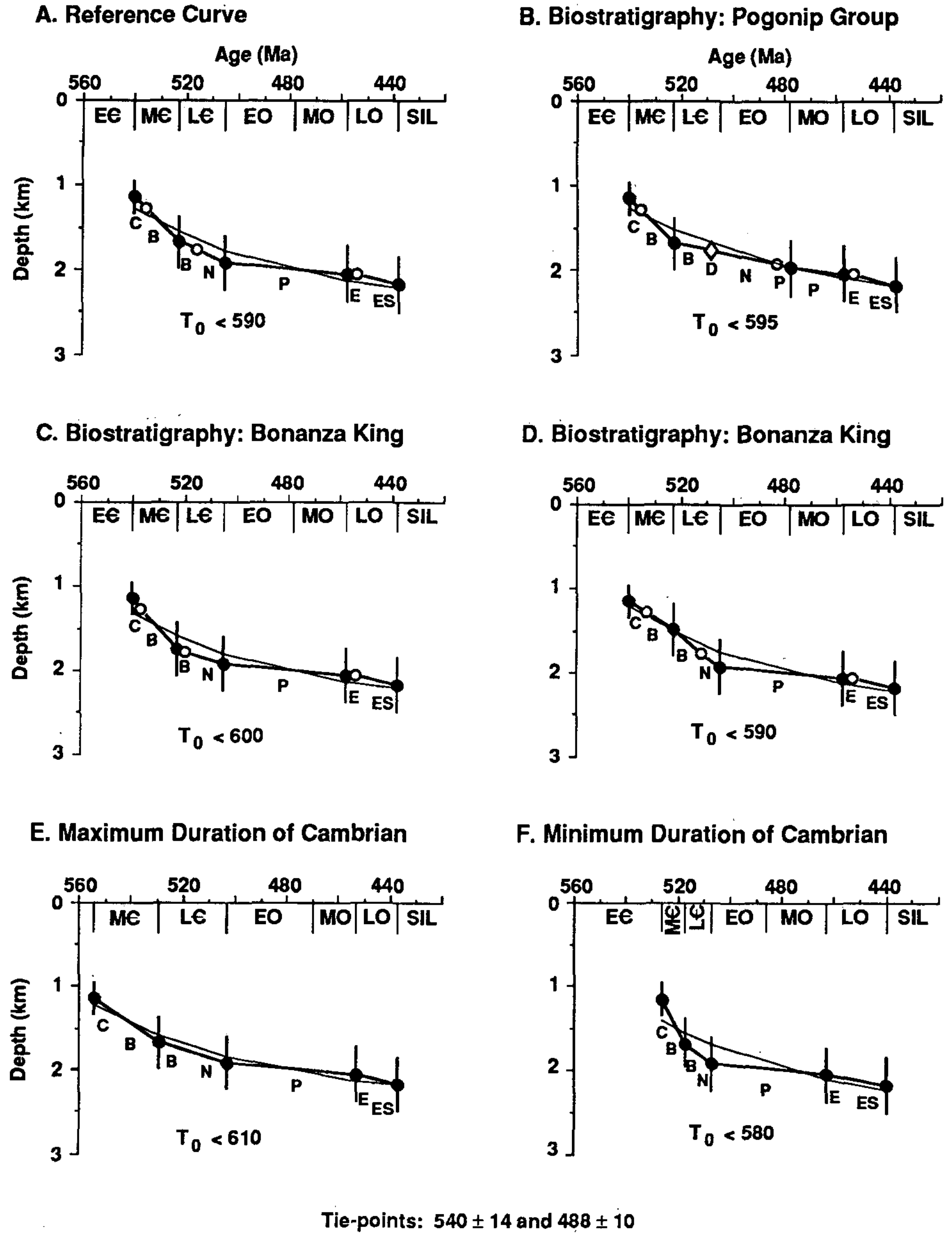

Figure 7. Sensitivity of the subsidence analysis to uncertainties in biostratigraphic age control and the Cambrian time scale. The tectonic subsidence curve from the Nopah Range (reference curve), using the most reliable stratigraphic thickness, biostratigraphic age control, absolute age control, and water depth estimates, is compared with modified tectonic subsidence curves when biostratigraphic age control and absolute age control are allowed to vary independently. In each graph, the heavy line is the R1 curve and the thinner line is the exponential curve. Solid circles represent formation boundaries tied to radiometric ages. Open circles represent formation boundaries that are interpolated linearly on the basis of present-day stratigraphic thickness. Vertical bars represent the range of tectonic subsidence based on maximum and minimum delithification values. A. Reference section. B. Biostratigraphic division in which the Early–Middle Ordovician boundary falls within unit 1 of the Pogonip Group (Miller, 1982). The age of the Dunderberg Shale, represented by the diamond, is interpolated and falls within the latest Cambrian, younger than suggested by its faunal assemblage. The estimate of T_0 is approximately the same as that of the reference curve. C. Minimum thickness of the Bonanza King (100 m) placed in the Late Cambrian. The estimate of T_0 is <600 Ma, slightly older than that of the reference curve. D. Maximum thickness of the Bonanza King (840 m) placed in the Late Cambrian. The estimate of T_0 is the same as that of the reference curve. E. The duration of the Cambrian Period is allowed to vary according to the error bars of the time scale of Harland and others (1982). In this case, the maximum duration is permitted, generating a curve that is roughly similar to the reference curve, suggesting a slightly older T_0. F. In this case, the duration of the Cambrian Period is allowed to vary to produce the minimum duration. The new curve generated suggests a younger T_0, with a slope that is considerably steeper than that of the reference curve.

Geologic Time Scale

Bond and Kominz (1984) and Bond and others (1988) compared several relatively similar time scales to determine their effect on subsidence analysis. Although absolute ages of individual boundaries from the time scales included in those studies vary somewhat from one time scale to another, differences in the duration of time-stratigraphic units are relatively small, and the use of any of these time scales has been found to produce curves with essentially the same form (Bond and Kominz, 1984; Bond and others, 1988). One exception to this is the time scale of Odin (Odin, 1982, 1985; Odin and others, 1983), in which the base of the Cambrian is very much younger (530 ± 10 Ma) than in other time scales, and the duration of the Cambrian much shorter (Fig. 8). Odin's time scale (1982) has been criticized by Harland (1983) with regard to its data base. Recent new data, however, lend support to a younger age for the Precambrian-Cambrian boundary, although at this time, an age of 530 Ma is not generally accepted (W. Compston and others, cited by Cowie and Johnson, 1985; Cope and Gibbons, 1987; S. M. Barr, 1988, personal commun.; Benus, 1988; Krogh and others, 1988; Conway Morris, 1988, 1989; Harland and others, 1989).

In this study, we use the time scale of Harland and others (1982; Fig. 8). The data base for the early Paleozoic portion of this time scale is drawn primarily from the Phanerozoic Time Scale (PTS; Harland and others, 1964), the Phanerozoic Time Scale Supplement (PTSS; Harland and

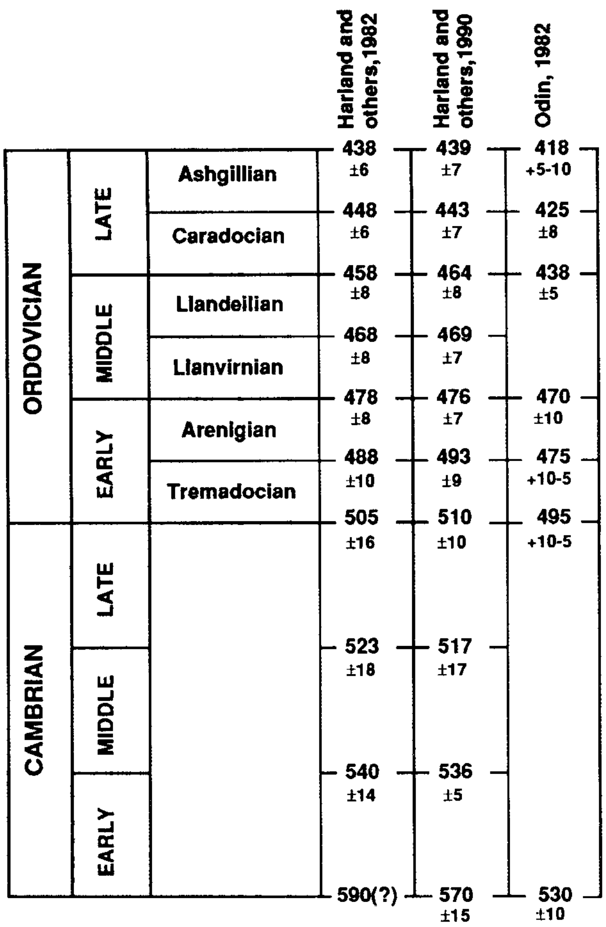

Figure 8. Comparison of the geologic time scale of Harland and others (1982), used in this study, with their revised time scale (Harland and others, 1990) and with that of Odin (1982) for the Cambrian and Ordovician Periods.

Francis, 1971), and the pre-Cenozoic data file of Armstrong (A; Armstrong, 1978). Numerical ages have been recalculated using the new decay constants (see Harland and others, 1982, p. 46). Radiometrically constrained tie points and interpolated ages are used to construct the time scale. For most stage boundaries older than 200 m.y., tie points are those ages that have an error range of less than 12 m.y. Exceptions to this are the poor tie points at the base of the Middle Cambrian (540 Ma) and at the base of the Arenigian in the Early Ordovician (488 Ma), which have error ranges of 28 and 20 m.y., respectively. Between these tie points, ages are interpolated linearly. On the basis of the midpoints and the maximum limits of these Cambrian and Ordovician tie points (that is, 540 ± 14 and 488 ± 10 Ma; Fig. 8; Harland and others, 1982), subsidence curves are constructed in which the duration of the Middle and Late Cambrian is permitted to vary (Figs. 7A, 7E, and 7F). In Figure 7E, the maximum duration of these time-stratigraphic units on the basis of the error ranges is assumed (51 m.y.), and the boundaries between the tie points are interpolated linearly, whereas in Figure 7F, the minimum duration is assumed (19 m.y.). In Figure 7A, the reference curve, the midpoint of the range of the tie points is used, yielding a 35-m.y. duration.

Because of the importance of the Middle Cambrian tie point in constructing the subsidence curves, we have examined in detail the data base and statistical approach used in the calculation of this age (Harland and others, 1982, p. 46–50). According to the methodology of Harland and others (1982), the age of a stage boundary is estimated and an error for that age is determined on the basis of the error values of individual determinations constraining the stage boundary. This procedure is repeated, assuming different estimates for the age of the stage boundary, until the error is minimized. Greatest importance is placed on those ages closest to the estimated age of a boundary, because only those ages from the older interval that are younger than the estimated age of the boundary and those ages from the younger interval that are older than the estimated age of the boundary constrain the age and error range of the boundary between the two intervals. A graphic representation of the data base for a given boundary is termed a "chronogram" (Fig. 9; Harland and others, 1982). We consider briefly all the age determinations in the data base for the Middle Cambrian tie point, but concentrate on those ages closest to the estimated boundary between the Early Cambrian and Middle Cambrian. There are five determinations in the data base for the Early Cambrian (Fig. 9). Of these, three are determinations on granite intrusions (PTSS353, PTS42, and A486 in Fig. 9) and are considered reliable. The remaining two determinations are K-Ar and Rb-Sr ages on glauconite (PTS185 and PTS183 in Fig. 9) and are suspect (Harland, 1983; Berggren and others, 1985; Obradovich, 1988). A K-Ar age on glauconite from the Kessyusse Beds in Siberia (536 ± 12 Ma; PTS185 in Fig. 9), however, is closest to the estimated boundary and is the key date from the Early Cambrian. For the Middle Cambrian, there are four determinations (Fig. 9). Of these, only one is a date on an igneous intrusion (Boisdale Hills granite, Cape Breton; PTS70 in Fig. 9). The remaining determinations are K-Ar and Rb-Sr ages

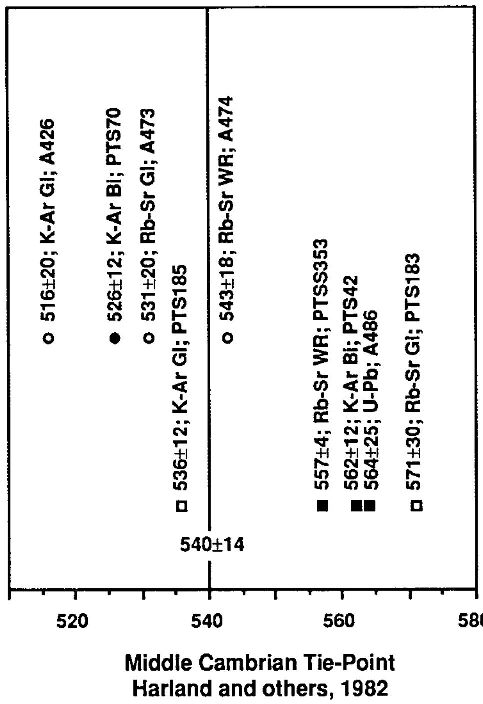

Figure 9. Chronogram for the Middle Cambrian tie point (modified from Harland and others, 1982). Circles are data points from the Middle Cambrian and younger; open circles are sedimentary ages on glauconite or shale; solid circles are igneous ages. Squares are data points from the Early Cambrian and older; open squares are sedimentary ages on glauconite; solid squares are igneous ages.

on glauconite or shale (A426, A473, and A474 in Fig. 9). Of these dates, all but the youngest age are close to the estimated age of the boundary. On the basis of these radiometric ages, Harland and others' statistical method predicts an error range of 28 m.y., with the Middle Cambrian tie point at 540 Ma. Although it is preferable to use igneous determinations for older rocks, it is not possible to eliminate all the sedimentary ages and retain enough data points to use the statistical approach of Harland and others (1982).

Re-evaluation of two important data points suggests that 540 Ma is a reasonably reliable age for the base of the Middle Cambrian. Recent new dating of granitoid rocks in the Boisdale Hills region of Cape Breton yields an older age of 574 ± 11 Ma (Poole, 1980; Barr and Setter, 1986). Re-examination of the field relations indicates that the Boisdale Hills granite does not intrude Cambrian sediments as previously thought, but is unconformably overlain by the Middle Cambrian Bourinot Group, thus suggesting that the stratigraphic age of the Boisdale Hills granite is Early Cambrian or older (Barr and Setter, 1986; S. M. Barr, 1988, personal commun.). In the revised data base for the Middle Cambrian tie point, the Boisdale Hills granite is placed in the Early Cambrian and no longer affects the estimated age of the boundary. Nonetheless, the best estimate for the base of the Middle Cambrian remains at 540 Ma.

A second potential data point thought to constrain the age of the base of the Middle Cambrian is the Ercall Granophyre in Shropshire, England. The nature of the contact between the Ercall Granophyre and the overlying Cambrian sediments has been uncertain, but recent excavation at the Ercall Quarry has exposed this contact. New structural and sedimentary evidence now indicates that the contact between the Ercall Granophyre and the Early Cambrian (late Tommotian) Wrekin Formation is an unconformity, with the Ercall Granophyre having intruded the Uriconian volcanic rocks (Precambrian), and subsequently having been uplifted and eroded prior to deposition of the Wrekin Formation (Cope and Gibbons, 1987). Thus, the stratigraphic age of the Ercall Granophyre is pre–late Tommotian. The age of the Ercall Granophyre has been determined by Patchett and others (1980) as 533 ± 13 Ma, using the Rb-Sr method. A widely circulated U-Pb age of 531 ± 5 Ma by ion microprobe on zircon from the Ercall Granophyre (W. Compston and others, cited by Cowie and Johnson, 1985) must be increased by 3.8% to 551 Ma, because of revision in the conventional U-Pb age of the ion-probe zircon standard (W. Compston, 1988, personal commun.). No further discussion of this data point seems warranted until the authors have published their primary data.

Although the age of the base of the Cambrian does not directly affect the Middle Cambrian tie point unless it is ≤540 Ma, some recently published data suggest that the age of this boundary may in fact be younger than previously thought. Strongest support of a younger age for the base of the Cambrian comes from the Avalon Peninsula in Newfoundland (Benus, 1988; Conway Morris, 1988, 1989). The best-constrained date is from a tuff in the Mistaken Point Formation. The tuff is interbedded with strata that contain Ediacaran fauna and has yielded a U-Pb zircon age of 565 ± 3 Ma (Benus, 1988). Approximately 8.7 km of sedimentary rock is present between this horizon and the first appearance of Early Cambrian (Tommotian) fauna (Benus, 1988). In the absence of undetected structural complexities, these data strongly indicate that the base of the Cambrian is younger than 565 Ma and perhaps closer to the age suggested by Odin (Conway Morris, 1988, 1989). The base of the Cambrian is given as 560 Ma in Figure 2.

In a revised version of their 1982 time scale, Harland and others (1990) considered the base of the Middle Cambrian to be slightly younger (536 Ma), primarily on the basis of the 531 ± 5 Ma age determination for the Ercall Granophyre in Shropshire, England, mentioned above (W. Compston and others, cited by Cowie and Johnson, 1985; Cope and Gibbons, 1987). If this point is eliminated from the data base, then the age of the base Middle Cambrian becomes 539 Ma, virtually identical to that of the 1982 time scale. On the basis of the revised time scale (assuming an age of 539 Ma for the base Middle Cambrian), the duration of the Middle and Late Cambrian is 29 m.y., which falls well within the error range for this same time interval in the 1982 time scale (see above).

The geologic time scale is subject to continual refinement as more data are incorporated and advanced techniques allow more accuracy and precision in age determinations. It is crucial to understand how future changes in a time scale will affect the results of this study. As mentioned above, the changes in the Middle and Late Cambrian part of the time scale have the most influence on our results. In general, a time scale with an increased duration of the Middle and Late Cambrian will produce a subsidence curve with a gentler slope and predict a slightly older estimate of T_0 (Fig. 7E). Significantly older estimates of T_0 can be obtained only if the base of the Middle Cambrian is significantly older than 540 Ma, which at present seems unlikely because recent estimates of the age of the Precambrian-Cambrian boundary indicate that it is younger than 565 Ma and possibly as young as 540 Ma (Benus, 1988; Conway Morris, 1988, 1989). A reduced duration of the Middle and Late Cambrian results in a steeper subsidence curve and predicts a slightly younger age of initiation of thermal subsidence (Fig. 7F). Changes in the duration of the Ordovician will have little effect on our results because by the Ordovician, most of the thermal anomaly had dissipated. If there were a systematic change toward younger ages for time boundaries, with the duration of time-stratigraphic intervals remaining constant, this would be reflected in a similar systematic shift to a younger estimate of T_0, but would not affect the stratigraphic position of the T_0 horizon. In general, however, reasonable variations in the time scale do not significantly affect estimates of T_0.

Paleobathymetry

In the construction of tectonic subsidence curves, the correction for paleo–water depths of <100 m is negligible (Bond and Kominz, 1984). In all the localities of this study, the facies are interpreted to represent shallow-water shelf deposition (Halley, 1974; Stewart and Poole, 1974; Kepper, 1976, 1981; Miller, 1976, 1982; Sloan, 1976; Palmer and Halley, 1979; Cooper and others, 1981, 1982; R. H. Miller and others, 1981; Sundberg, 1982). An association of features, including fenestral fabric, microbial laminae, burrow mottling, and desiccation cracks, suggests shallow-water deposition. Fossil assemblages, consisting of trilobites, brachiopods, pelmatozoans, stromatolites, and thrombolites, also support a shallow-water subtidal to intertidal shelf interpretation. Therefore, it is unnecessary to include a water-depth correction, because changes in water depths are small in comparison with the total amount of subsidence.

DISCUSSION

Geologic Evidence for Rifting in Late Proterozoic and Cambrian Time

Several stratigraphic units considered to be related to rifting have previously been identified and described in the Middle Proterozoic through Lower Cambrian rocks of western North America, with the most convincing geologic evidence for rifting present in rocks older than about 700 Ma. In the Mackenzie Mountains of northwestern Canada, for example, rifting is well documented in strata of Late Proterozoic age immediately underlying the glacial beds and appears to have ended during deposition of post-glacial siltstones (Jefferson, 1978; Eisbacher, 1981, 1985). In eastern California, rifting associated with the Amargosa basin ended during deposition of the Noonday Dolomite (older than 700–680 Ma; Wright and others, 1976; Roberts, 1982; Miller, 1985). A number of geologic observations, however, suggest that crustal extension may have continued into Cambrian time, corroborating the timing indicated by subsidence analysis (Christie-Blick, 1984; Bond and others, 1985; Christie-Blick and Levy, 1989a). The most important of these observations is the presence through much of the western United States of volcanic rocks as young as Early Cambrian age (Christie-Blick and Levy, 1989a). Although volumetrically limited, the rocks are strongly suggestive of continued extension, because for the most part igneous activity either predates or coincides with times of crustal extension and is unusual in the post-rift phase of passive continental margins (for example, Bally and others, 1981; Scrutton, 1982; Watkins and Drake, 1982). Other evidence for tectonic activity includes local angular unconformities, facies evidence for marked changes in either the orientation or steepness of the paleoslope, and compositional variations in sandstones that are consistent with local uplift and erosion of crystalline basement (Christie-Blick, 1984; Bond and others, 1985; Christie-Blick and Levy, 1989a).

Volcanic rocks are most abundant at the stratigraphic level of the diamictites (Crittenden and others, 1983; Link, 1983; Christie-Blick, 1985; Miller, 1985; Harper and Link, 1986), but they are also known in Nevada from rocks as young as the uppermost Stirling Quartzite (Early Cambrian; Stewart, 1974), and in Utah from the Browns Hole Formation, lower Prospect Mountain Quartzite, and lower Tintic Quartzite (latest Proterozoic to Early Cambrian; Morris and Lovering, 1961; Crittenden and others, 1971; Crittenden and Wallace, 1973; Abbott and others, 1983). The rocks consist of mafic to intermediate-composition flows and sills, and lesser amounts of volcaniclastic sandstone, conglomerate and breccia, and tuffaceous shale. Available chemical data suggest that the volcanic rocks in the lower part of the section are of tholeiitic to alkalic affinity (Stewart, 1972; Harper and Link, 1986), consistent with the range of compositions for igneous rocks in extensional continental settings (Barberi and others, 1982). In northern Utah and southeastern Idaho, for example, mafic volcanic rocks interstratified with diamictite are high-Ti, high-Zr/Y within-plate basalts, with Nb/Y ratios and patterns of light-REE enrichment indicative of transitional tholeiitic-alkalic to alkalic compositions (Harper and Link, 1986). Similar results were obtained by Devlin and others (1985) for mafic volcanic rocks from the approximately correlative Huckleberry Formation of northeastern Washington, and by Reid and Sevigny (1988) and Sevigny (1988) from metavolcanic rocks of the Horsethief Creek Group in southwestern Canada. Comparable data have not yet been obtained for the younger volcanic rocks in the western United States, although the Browns Hole Formation appears to contain rocks of alkalic affinity (Crittenden and Wallace, 1973).

Angular unconformities are unusual in the post-glacial section of the western United States (Stewart, 1970; Crittenden and others, 1971), but a good example with angular discordance of about 10° is present in the Wasatch Range southeast of Salt Lake City, Utah, at the base of the Tintic Quartzite (Early to Middle Cambrian). Less impressive unconformities have been mapped in eastern California at the base of both the middle member of the Wood Canyon Formation (Early Cambrian) and the Stirling Quartzite (latest Proterozoic to Early Cambrian). All of these unconformities are consistent with local deformation of the crust, although only the Tintic unconformity requires structural tilting (Christie-Blick and Levy, 1989a). Considerably more prominent angular unconformities are present in the Canadian Cordillera beneath the Gog Group and Backbone Ranges Formation (both latest Proterozoic to Cambrian; Hofmann and Aitken, 1979; Eisbacher, 1981; Bond and others, 1985; Aitken, 1989). The Canadian examples indicate a major tectonic event of regional extent at about the inferred time of onset of thermal subsidence.

Local changes in the orientation or steepness of the paleoslope also are suggestive of crustal deformation, perhaps related to continuing extension (Christie-Blick and Levy, 1989a). In eastern California, the Wood Canyon Formation records an abrupt change in paleoslope orientation (Diehl, 1976, 1979). The middle member consists of as much as several hundred meters of conglomeratic and arkosic sandstone, interpreted by Diehl to have accumulated in a braided alluvial to tidally dominated marine environment. The rocks thicken and coarsen to the northeast and yield southwest-directed paleocurrents. In comparison, overlying and underlying stratigraphic units thicken to the west or northwest and yield west- to north-directed and polymodal paleocurrents. Diehl (1976, 1979) suggested that the middle part of the formation accumulated in a fault-bounded basin and that the immature sediment was derived from an uplifted source area northeast of the Death Valley region.

Most of the Late Proterozoic to Cambrian sandstones of the western United States are exceedingly mature both texturally and compositionally (Christie-Blick and Levy, 1989a, 1989b). The volume of sediment involved precludes significant recycling from older sandstones and suggests derivation directly from crystalline basement either by intense chemical weathering in a warm, humid climate (Chandler, 1988; Johnsson and others, 1988; Christie-Blick and Levy, 1989b) or by eolian deflation and preferential removal of feldspar and lithic fragments as wind-blown dust (Dalrymple and others, 1985; R. W. Dalrymple, 1989, personal commun.). The local presence of immature feldspathic sandstones in the Wood Canyon Formation as well as in the Mutual Formation and Geertsen Canyon Quartzite of Utah indicates temporary variations in the efficiency of weathering processes, perhaps as a result of crustal uplift and enhanced rates of erosion.

Isotopic dating may provide an independent means of comparing the inferred age of onset of thermal subsidence determined from the subsidence analysis with the geologic evidence for latest Proterozoic and Early Cambrian rifting in the southern Great Basin. A tuffaceous bed near the base of the Johnnie Formation in the Nopah Range has yielded zircons amenable to U-Pb dating (A. P. LeHuray, 1990, personal commun.). If the age of this tuff is >590 Ma, it can be assumed that extension continued after deposition of the Noonday Dolomite and that onset of thermal subsidence in latest Proterozoic or Early Cambrian time corresponds to a horizon stratigraphically as high as the Stirling Quartzite or Wood Canyon Formation (Fig. 2). If the age of the tuff is <590 Ma, we cannot eliminate the possibility that post-rift thermal subsidence began during deposition of the Johnnie Formation. In that case, however, either the ages inferred from the stromatolites within the Johnnie Formation and the Noonday Dolomite are incorrect or a significant hiatus is present in the transition between these units. On the basis of results presented in this paper and available geologic data, we predict that the tuff is about 650 Ma (or older) and that the Amargosa basin is unrelated to the phase of thermal subsidence associated with the early Paleozoic passive margin.

Sea Level

In all of the subsidence curves generated for the early Paleozoic passive continental margins in both eastern and western North America, the exponential curve deviates from the R1 curve for Middle and Late Cambrian time probably as a result of eustatic change (Bond and others, 1988; this study). Bond and others (1988) have shown that a measure of eustasy can be recovered from the R1 curve by removing the tectonic subsidence, assumed to correspond approximately with the best-fit exponential having a decay constant of 62.8 m.y. The deviation of the R1 curve from the exponential is plotted as apparent sea level versus age, and is called the second reduction of the data or the R2 curve (Bond and others, 1988).

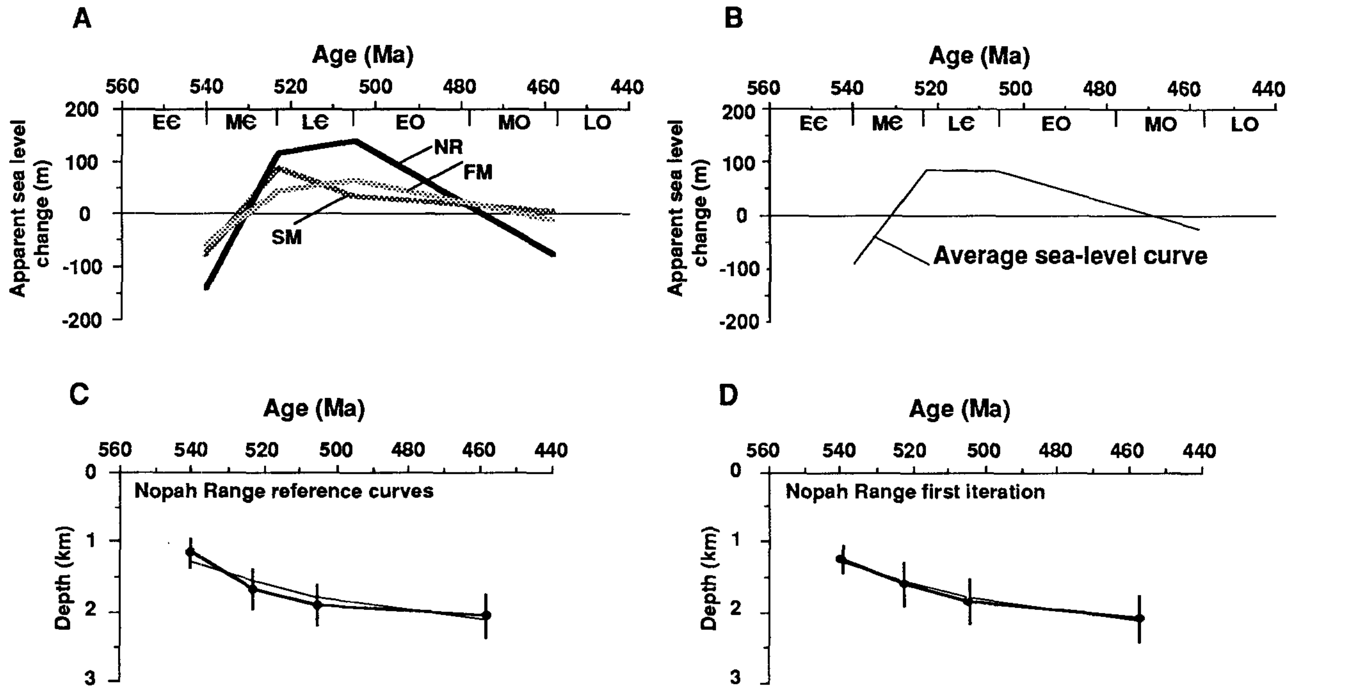

Figure 10. A. Plot of the R2 curves for the Nopah Range (NR), Spring Mountains (SM), and Funeral Mountains (FM). The Inyo Range is not included because the strata have experienced structural thinning. The R2 curve is determined by subtracting the exponential curve from the R1 curve and plotting the deviation. Apparent sea-level change is the difference between the R1 curve and the exponential curve (Bond and others, 1988). The zero datum differs from present-day sea level by an unknown amount. The R2 curve indicates the trend of long-term eustasy and is an approximation of the magnitude (see text). A sea-level rise is indicated in the Middle to Late Cambrian, followed by a sea-level fall in the Early Ordovician. B. Iterative approach of Watts and Steckler (1979) of isolating eustatic and tectonic effects in subsidence analysis. An average sea-level curve is generated by averaging R2 curves shown in Figure 10A. C. Nopah Range reference curve. Heavy curve represents the R1 curve; thinner curve is the best-fit exponential curve. D. Result of the first iteration of averaging the R2 curves and removing that curve from the original R1 curve. Heavy curve is the new R1 curve generated; thinner curve is the best-fit exponential with a decay constant of 62.8 m.y. to the new R1 curve.

The R2 curve is only an approximation of eustasy for several reasons. (1) The zero datum, taken as the point at which the R1 curve and the exponential curve coincide, differs from present-day sea level by an unknown amount (Bond and others, 1988). (2) The best-fit exponential is only an approximation for tectonic subsidence because the eustatic component of the R1 curve clearly varies through time. (3) The value of the decay constant of the exponential may deviate from 62.8 m.y. (Bond and others, 1988; Marty and Cazenave, 1989). (4) The lithosphere is assumed to have no strength, and thus changes in water loading are compensated only locally (Airy compensation). (5) The R2 curve contains no correction for the effects of water loading, which for local isostatic compensation increases the magnitude of apparent sea-level change by a factor of about 1.4.

The R2 curves generated for the California margin indicate a eustatic rise in the Cambrian, with a maximum rise occurring between 525 and 505 Ma, followed by a eustatic fall in the Ordovician (Fig. 10A). The timing and approximate magnitude of the eustatic fluctuation agree well with those inferred for western Canada, Utah, and the southern Appalachians (Bond and others, 1988). A correction for the effects of water loading yields an apparent eustatic rise of about 120 m in the Middle to Late Cambrian, similar to estimates of the magnitude of eustatic rise suggested for the Mesozoic (Kominz, 1984; Harrison, 1990).

Watts and Steckler (1979) suggested that the effects of eustasy can be separated from tectonic subsidence, represented by that part of the curve that is exponential in form, by employing an iterative process. The deviations of the R1 curve from the exponential curve for each locality studied, with the exception of the Inyo Range, are averaged together, following their procedure, to generate an average sea-level curve (Fig. 10B). This procedure tends to remove local effects and generates an estimate of the eustatic signal that is uniform from one locality to another. This average sea-level curve, which may be a more accurate estimate of the magnitude of fluctuation, is then removed from the R1 curve, resulting in a new tectonic subsidence curve. A new best-fit exponential with a decay constant of 62.8 m.y. is fit to this tectonic subsidence curve and the process is repeated until the deviation between the tectonic subsidence curve and the exponential is minimized. The result of the first iteration of this process for the Nopah Range is compared with the reference curve (Figs. 10C and 10D). Removing the effects of eustatic sea level markedly improves the fit of the tectonic subsidence curve to the exponential curve; however, slight deviations still remain. These deviations cannot be explained by eustatic sea-level changes, and may be the result of local tectonic effects, uncertainties in biostratigraphic age control, changes in the rate of sedimentation and/or breaks in sedimentation, or minor structural complications affecting stratigraphic thicknesses.

Comparison with Previous Studies

Results from this study are in general agreement with the conclusions of previous studies of subsidence across the early Paleozoic passive margin of western North America (Stewart and Suczek, 1977; Armin and Mayer, 1983; Bond and others, 1983; Bond and Kominz, 1984), which demonstrated that thermal subsidence in the Cordilleran miogeocline began between 600 and 550 Ma. Of these studies, those by Bond and others (1983) and Bond and Kominz (1984) are definitive, but refer to areas of the western United States and Canada north of the localities discussed in this paper. In an earlier study, Stewart and Suczek (1977) suggested that in eastern California, thermal subsidence may have begun between 650 and 600 Ma, but their conclusions are based on uncorrected stratigraphic thickness and in part on the stratigraphic position of the Precambrian-Cambrian boundary, which is not well constrained.

Armin and Mayer (1983) analyzed the tectonic subsidence at four localities in the western United States, including the Panamint Range, California (PR in Fig. 1; Fig. 11A), but their results are difficult to evaluate. In correcting for the effects of compaction in fully lithified rocks, stratigraphic thickness and decompacted thickness should be the same only for the time of deposition. For the Panamint Range section, however, Armin and Mayer showed these two curves converging in the late Paleozoic (Fig. 11A). Furthermore, unlike the subsidence curves presented in this paper, the curves for the Panamint Range are remarkably linear after Early Cambrian time (Fig. 11B) and therefore not interpretable according to the criteria set out above. The curvature shown depends strongly on the assumed position of the Proterozoic-Cambrian boundary.

Implications for Basin Evolution

Geologic observations and isotopic dating indicate that at least two discrete rifting events occurred in the western United States during Late Proterozoic and Early Cambrian time, one at ~800 to 700 Ma (Stewart, 1972, 1976; Armstrong and others, 1982; Evenchick and others, 1984; Devlin and others, 1985, 1988), and another at ~590 Ma (Christie-Blick, 1984; Bond and others, 1985). The earlier event appears to be too early to be reconciled with the results of quantitative subsidence analysis because only a residual thermal anomaly would remain from this event at 590 Ma. On the other hand, the apparent discrepancy between the small amount of observed upper-crustal extension associated with the latest Proterozoic and Early Cambrian event and the relatively rapid thermally driven subsidence characterizing the later Cambrian and Ordovician needs to be resolved. Extension of the upper crust during latest Proterozoic time may have been only a few percent because greater amounts of extension commonly result in significant tilting of fault blocks and in the development of angular unconformities (for example, Stewart, 1980; Barton and Wood, 1984; Gibbs, 1984). In contrast, subsidence analysis suggests that the lithosphere as a whole was extended by at least several tens of percent and possibly as much as 100% (apparent β factors close to 2 in Fig. 5), although in detail such estimates are model dependent and require assumptions about the mechanism and duration of extension, the time of onset of thermal subsidence, and the flexural rigidity of the lithosphere.

Three hypotheses are suggested to reconcile this apparent inconsistency between regional lithospheric thinning and the absence of appreciable evidence for upper crustal extension after deposition of the Noonday Dolomite. One idea is that the lack of evidence for crustal extension is due to limited exposure of rocks of Late Proterozoic and Early Cambrian age (Levy and Christie-Blick, 1989). Existing exposures tend to be aligned in a north-south direction, approximately parallel to the expected preferred orientation of Late Proterozoic normal faults, and outcrops of specific units tend to be small in comparison with the typical spacing of major normal faults in extensional terranes (2–40 km; Hamilton and Myers, 1966; de Charpal and others, 1978; Stewart, 1978; Hauge and others, 1987; Rosendahl, 1987). The existence of Late Proterozoic faults is also likely to be obscured by Mesozoic and Cenozoic deformation. Normal faulting is commonly accompanied by the tilting of fault blocks and by the local development of angular unconformities at the contact between pre-rift and syn-rift deposits, within the syn-rift section, and between syn-rift and post-rift strata (Evans and Parkinson, 1983; Harding, 1984; Hutchinson and others, 1986). The lack of obvious discordance within the Late Proterozoic and Early Cambrian stratigraphy at most localities in the western United States, however, may also be due in part to limited exposure.

A second explanation for the lack of evidence for significant crustal extension after deposition of the Noonday Dolomite is that the lithosphere may have extended in a heterogeneous manner, possibly in association with regional detachment faults. Detachment models, in which extension in the upper and lower lithosphere is geographically partitioned, permit the development of considerable post-rift subsidence even in areas for which there is little evidence for extension within the upper crust (Wernicke, 1985; Lister and others, 1986; Kusznir and others, 1987; Buck and others, 1988; Lister and Davis, 1989). For sections to balance on a regional scale, equivalent amounts of extension must be accommodated at all structural levels. This can be achieved in the western United States if relatively large amounts of extension in the lower crust and upper mantle beneath Utah and Idaho were balanced at an upper-crustal level in central and western Nevada near the Paleozoic edge of the continent, where evidence for

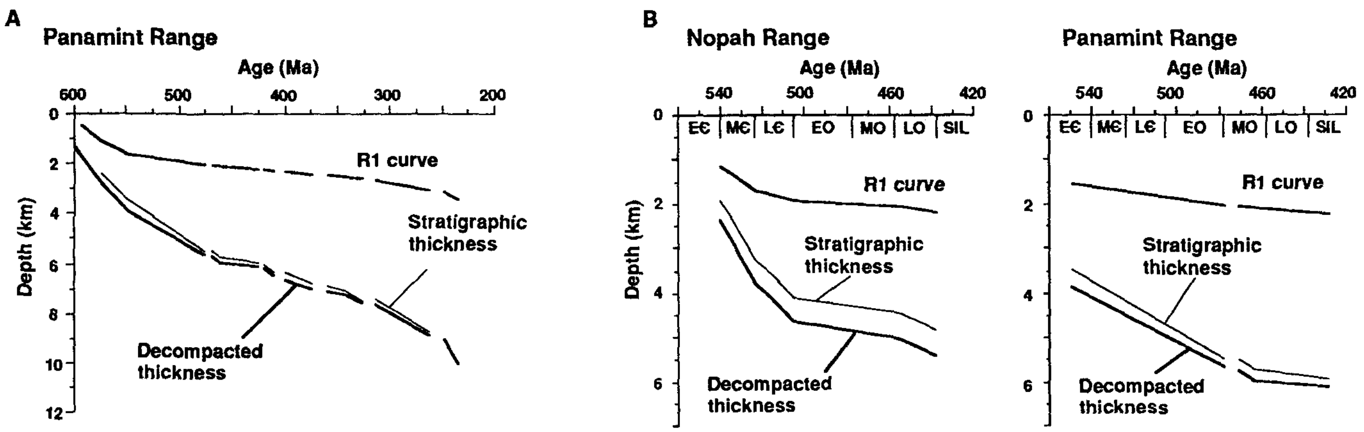

Figure 11. A. Tectonic subsidence curve from the Panamint Range (replotted from Armin and Mayer, 1983). Breaks in the curves represent unconformities in the stratigraphic record. B. Comparison of tectonic subsidence curves from the Nopah Range (this study) and the Panamint Range (Armin and Mayer, 1983; replotted at the same scale as the Nopah Range) from the base of the Middle Cambrian to the Silurian. Note the lack of any curvature in the R1 curve from the Panamint Range.

extension was either removed by continental separation in Early Cambrian time or hidden beneath the younger Paleozoic orogens (Christie-Blick and Levy, 1988, 1989a). The effect of a detachment fault soling below the crust is to produce significant uplift above areas of deep extension through the replacement of cooler lithospheric mantle by warmer lithospheric mantle (Kusznir and others, 1987). Since physical stratigraphic observations indicate only minimal uplift in the western United States during Late Proterozoic to Early Cambrian time, any detachment or detachments would have dipped to the east and soled into the lower crust (Christie-Blick and Levy, 1989a).

A third hypothesis is that some or all of the localities studied are located continentward of the hinge zone between stretched and unstretched lithosphere and that observed post-rift subsidence may be due in part to flexural loading by sediments that accumulated basinward of the hinge zone. A possible test of this hypothesis is to obtain more reliable stratigraphic data basinward of the Funeral, Nopah, and Spring Mountains (Fig. 3). For example, if the true stratigraphic thickness in the Inyo Mountains is significantly greater than that preserved, it may be possible to show that this locality is not within the flexural wedge. On the other hand, if stratigraphic thicknesses do not increase laterally toward the edge of the continent, it may be possible to argue that none of the localities studied is within the flexural wedge. In either case, the estimated maximum age of onset of thermal subsidence would still be valid. In fact, the estimated age of onset of thermal subsidence is likely to be overestimated because thicker stratigraphic sections imply more rapid subsidence, and hence a better-constrained T_0.

Ongoing research is aimed at testing these ideas. In spite of analytical difficulties and small samples, we hope that eventually it will be possible to calibrate the Proterozoic stratigraphic section by means of U-Pb geochronology. In addition, subsidence analysis will be undertaken in two dimensions along palinspastically restored transects (based on Levy and Christie-Blick, 1989) to assess the role of flexure and to obtain better constraints on mechanisms of extension.

CONCLUSIONS

Tectonic subsidence analysis of the Cambrian and Ordovician strata across the miogeocline in eastern California and southern Nevada indicates that subsidence decayed exponentially, consistent with a thermal mechanism of subsidence. The age of initiation of thermally driven subsidence, as determined by the analysis, is between 590 and 545 Ma. These results are based on the geologic time scale of Harland and others (1982) and are in agreement with studies from elsewhere in the Cordillera (Bond and others, 1983, 1985; Bond and Kominz, 1984).

We have evaluated potential sources of error in the analytical procedures and have found that predictions of the age of onset of thermal subsidence are sound in view of the range of our geologic assumptions. The sensitivity of the subsidence analysis to the Cambrian-Ordovician time scale was assessed and the predictions of the subsidence analysis were found to vary in a predictable manner with respect to the continual fine-tuning of the time scale. Nonetheless, likely uncertainties in the time scale do not significantly affect the estimate of the timing of onset of thermal subsidence.

Although the most convincing geologic evidence for crustal extension is in rocks older than ~700 Ma, a number of geologic observations suggest that crustal extension may have continued into Cambrian time, as indicated by the subsidence analysis. Three hypotheses have been presented to reconcile the apparent discrepancy between the lack of appreciable evidence for upper-crustal extension in latest Proterozoic and Early Cambrian time and the rapid thermal subsidence characterizing the early Paleozoic. (1) Much of the critical evidence for upper-crustal extension is not preserved in the limited exposures of latest Proterozoic and Lower Cambrian strata. (2) The lithosphere may have been thinned regionally after deposition of the Noonday Dolomite in association with one or more east-dipping detachments. (3) Some or all of the localities are continentward of the hinge zone, and the observed subsidence is due to flexural loading in an adjacent basin to the west.

A final result from the tectonic subsidence analysis is that the long-term Cambrian-Ordovician eustatic signal evident in other parts of North America is also present in our data. A misfit between the R1 curves and the exponential curves, consistent with a relative highstand in the Late Cambrian to Early Ordovician, cannot be eliminated from the results by including likely errors in stratigraphic thickness and age control.

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