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Tectonic controls of foreland basin subsidence and Laramide style deformation, western United States

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ABSTRACT

A variety of Late Mesozoic to Early Cenozoic tectonic events in the Rocky Mountains region are temporally and spatially coincident with inferred variations in kinematics of plate interactions and subduction geometries. This coincidence suggests that these disparate events are the products of a single causal mechanism. The following tectonic features are regarded as genetic expressions of variations in subduction modes and geometries: (1) the history of igneous activity in the western United States; (2) the contrasting styles and loci of deformation along the foreland fold and thrust belt (Sevier style) and the basement-cored uplifts (Laramide style) bordering the northern and eastern margins of the Colorado Plateau; (3) the development and maintenance of the Colorado Plateau as a relatively rigid tectonic block; (4) the timing and geometry of subsidence in the foreland basin; and (5) the disjunct history of subsidence and subsequent uplift of the Colorado–Wyoming region beyond the foreland basin.

During a period of normal subduction (from before 92 to about 80 Ma), thin-skinned décollement style deformation occurred along the Sevier fold and thrust belt opposite the convergent margin. Coeval subsidence of the foreland basin was confined to a relatively narrow zone to the east of the Sevier belt. This subsidence is attributed to lithospheric flexure induced by supracrustal loading of thrust plates and sediment. Geohistory analyses of strata along the axis of the foreland basin indicate that foreland basin subsidence began as early as 115 Ma, with a major episode of rapid subsidence initiated by about 90 Ma.

During an ensuing period of low-angle subduction, the Colorado Plateau was underpinned by subducted lithosphere and behaved as a mechanically rigid block, a consequence of the doubled lithospheric thickness. From about 80 to 67 Ma, rapid subsidence occurred over an anomalously broad region centred about Colorado and Wyoming. This episode of subsidence is attributed to sublithospheric loading and cooling induced by the shallowly subducted oceanic plate. To the north and south of the Colorado–Wyoming locus, foreland basin subsidence continued without interruption coincident with continued foreland folding and thrusting. Another effect of low-angle subduction was the transfer of deformation from the Sevier belt (termination about 75 Ma) to the eastern and northern margins of the Colorado Plateau, coincident with the position of greatest contrast in mechanical properties of the lithosphere. This initiated Laramide style basement-cored uplifts at about 69 Ma. Decoupling of subducted lithosphere from overlying lithosphere at about 50 Ma caused regional uplift and erosional stripping of the Colorado–Wyoming region, lithospheric flexure to the east, and sediment accumulation on the High Plains following a long period of non-deposition.

INTRODUCTION

The Late Mesozoic foreland basin of the western United States displays an unusual discordance in timing, areal limits and geometry of subsidence. In the first phase of its development, the foreland basin existed as a north–south trending, relatively narrow, asymmetric structural trough adjacent and parallel to an eastward advancing foreland fold and thrust belt.

This simple pattern of subsidence adjacent to a linear foreland fold and thrust belt was modified during the Late Cretaceous with the development of a second mode of subsidence in Colorado and Wyoming. In that area subsidence occurred in a broad, sub-circular region centred well to the east of the fold and thrust belt. This second mode of subsidence was coeval with continued development of the foreland fold and thrust belt/foreland basin pair to the north and south of the Colorado–Wyoming region.

In the quest for an explanation of these two spatially disjunct modes of foreland basin subsidence, a coincidence was recognized in the temporal and spatial occurrences of several other seemingly unrelated tectonic events and elements. This coincidence invites an explanation that unites these disparate elements and events by a single causal mechanism. In this study, a large number and variety of geological data were assembled, evaluated and integrated. These constituted the basis for an analysis of the spatial and temporal occurrences and the causes of Late Mesozoic through Early Cenozoic deformation in the western United States. This report presents a synopsis of these data, reviews their significance, and proposes a unified explanation for the genesis of several major tectonic features and events in the Rocky Mountains region.

During the Late Mesozoic and Early Cenozoic, oceanic plates of the Pacific Ocean were converging with and subducting beneath the western margin of North America. Calculated and inferred reconstructions of plate motions indicate that kinematics of subduction and geometries of subducted oceanic plates varied in time and space. Several published models relate absolute plate motions, kinematics of plate interactions, and varying modes of subduction to states of stress, styles and positions of deformation, and emplacement of magmas within the overriding continental plate (e.g. Coney, 1972; Sykes, 1978; Dewey, 1980; Zoback & Zoback, 1980; Cross & Pilger, 1982; Engebretson, Cox & Thompson, 1984, among others). With a knowledge of plate interactions, such models offer predictions about the nature and occurrence of deformation and magmatism which may be tested against geological observations. Alternatively, in the absence of reliable plate reconstructions, they provide a basis for interpreting the causes of observed deformational and magmatic events.

Objectives and outline

The primary intention of this report is to demonstrate temporal and spatial coincidence between the history of plate interactions and the occurrence of major deformational and magmatic events in the western United States. The second intention is to propose a unified explanation for these events which relates them genetically to the kinematics and dynamics of plate interactions and the mechanical properties of the lithosphere within which they occurred. The occurrence and genesis of the following tectonic elements and events are specifically addressed:

(1) the history of igneous activity in the western United States;
(2) the contrasting styles and loci of deformation along the foreland fold and thrust belt (Sevier style) and the basement-cored uplifts (Laramide style) bordering the northern and eastern margins of the Colorado Plateau;
(3) the development and maintenance of the Colorado Plateau as a relatively rigid tectonic block;
(4) the timing and geometry of subsidence in the foreland basin; and,
(5) the disjunct history of subsidence and subsequent uplift of the Colorado–Wyoming region beyond the foreland basin.

Some of the data and interpretations presented in this report have been published previously. Nonetheless, they are summarized briefly in order to provide an integrated and coherent assessment of the spatial and temporal association among some of the tectonic events and elements listed above. Following the summaries of previous work, new information and compilations of data are presented. These augment the range of tectonic elements and events in the western United States that are temporally and spatially associated. This documentation of coincidence in timing and spatial occurrence among these events provides the basis for suggesting a genetic relationship and for proposing a mechanism that accounts for the observed associations.

Published plate reconstructions provide the temporal, kinematic and dynamic frameworks with which deformational and magmatic events in the western United States may be compared and related. These reconstructions and their geological implications are discussed first. Next, the history of Late Mesozoic to Early Cenozoic magmatism in the western United States is reviewed. Because this history has been thoroughly documented in a number of previous reports, it is summarized cursorily in order to demonstrate consistency and probable genetic relations between the nature of plate interactions and resultant igneous activity.

The history of deformation in the Rocky Mountains region is considered next. This discussion focuses on the disjunct modes of foreland basin subsidence, on the development and maintenance of the Colorado Plateau as a rigid tectonic block, and on the contrasting styles and loci of deformation along the fold and thust belt (Sevier style) and the basement-cored uplifts

(Laramide style) bordering the northern and eastern margins of the Colorado Plateau. In contrast with the excellent summaries of magmatic history, there does not exist a comparable synthesis of deformation in the Rocky Mountains region for the same time period. Documentation of the varying styles and histories of deformation are contained either in studies of specific, local structural elements or in reviews more limited in regional and temporal scope than is attempted in this report. From a critical review of pertinent literature, a set of geological data that constrains the temporal and spatial limits of deformation in the Rocky Mountains region was assembled. Only those tectonic elements for which geological evidence provides an accurate age of initiation, duration, and/or cessation of deformation are included in this synopsis. These data demonstrate spatial and temporal coincidence between specific modes and geometries of subduction and the major tectonic events and elements listed above.

Finally, a genetic relationship is proposed among the several, seemingly independent, tectonic events and elements and the history of plate interactions along the western United States. From theoretical considerations and empirical observations of contemporary subduction systems, the geological expressions of particular modes of subduction have been recognized (as summarized by Cross & Pilger, 1982, for example). These provide a rationale for explaining the observed histories of deformation and magmatism in the western United States by a particular evolution of plate interactions. Moreover, the changes in geometry of subducted lithosphere associated with this evolution provide a mechanism that accounts for the temporal and spatial occurrence and the nature of deformation in the Rocky Mountains region. It appears that several major, seemingly unrelated, tectonic events and elements in the western United States are the united consequences of a particular evolution of plate interactions.

HISTORY OF PLATE INTERACTIONS FROM PLATE RECONSTRUCTIONS

The history of interaction between North America and the oceanic Kula, Farallon and Pacific plates originally was described by Atwater (1970) and Atwater & Molnar (1973). This history, as determined by rotations of magnetic signatures of oceanic crust, is reasonably well constrained for the past 80 Myr. Refinements and temporal extensions of their plate reconstructions have been accomplished by inclusion of the hot spot, or absolute motion, reference frame in reports by Engebretson, *et al.* (1984), Engebretson, Cox & Gordon (1984), Henderson, Gordon & Engebretson (1984) and Jurdy (1984), among others. Absolute motion models extend the reconstructions to 150 Ma. Although these are non-unique solutions, their validity may be assessed by comparison with the temporal and spatial occurrence of geological events in the western United States that reflect plate interactions. Plate reconstructions for times earlier than 150 Ma rely on palaeomagnetic data from continents and uniformitarian arguments relating varying modes of subduction to observed temporal and spatial patterns of magmatism and deformation in the western United States.

The absolute motion of North America and the relative convergence between North America and the Farallon plate, as derived by Engebretson and co-workers, are shown schematically in Figs 1 and 2. More graphic, but inferential, representations of the history of plate interactions and variations in subduction geometries are presented in Figs 8 and 9. From about 135 to 127 Ma, the Farallon plate was converging with North America at a rate of about 70 km Ma^{-1}, and from 127 to 100 Ma the convergence velocity decreased slightly to about 55 km Ma^{-1}. From 100 to 75 Ma, convergence was oblique to the subduction zone, but at an increased velocity of about 100 km Ma^{-1}. During the entire 145–85 Ma period, North America was moving to the NW at velocities of about 30 km Ma^{-1} in the hot spot reference frame, or oblique to the inferred N30°W trend of the subduction zone separating the North American and Farallon plates. These rates and orientations of relative convergence combined with the oblique absolute motion of North America toward the trench should result in normal (moderate- to steep-angle) subduction and development of a volcanoplutonic arc along the western margin of North America. Moderate-angle subduction also should be espressed as Cordilleran style crustal shortening and foreland fold and thrust deformation as the hot, ductile, isostatically uplifted back-arc region is compressed against the colder, more rigid craton (Armstrong & Dick, 1974; Cross & Pilger, 1982).

At about 75 Ma, the North American plate changed direction in the absolute motion frame and moved south-westward, toward and normal to the trench. Between about 65 and 47 Ma, the absolute motion of North America attained a maximum rate of nearly 50 km Ma^{-1}. From 75 to 44 Ma the relative conver-

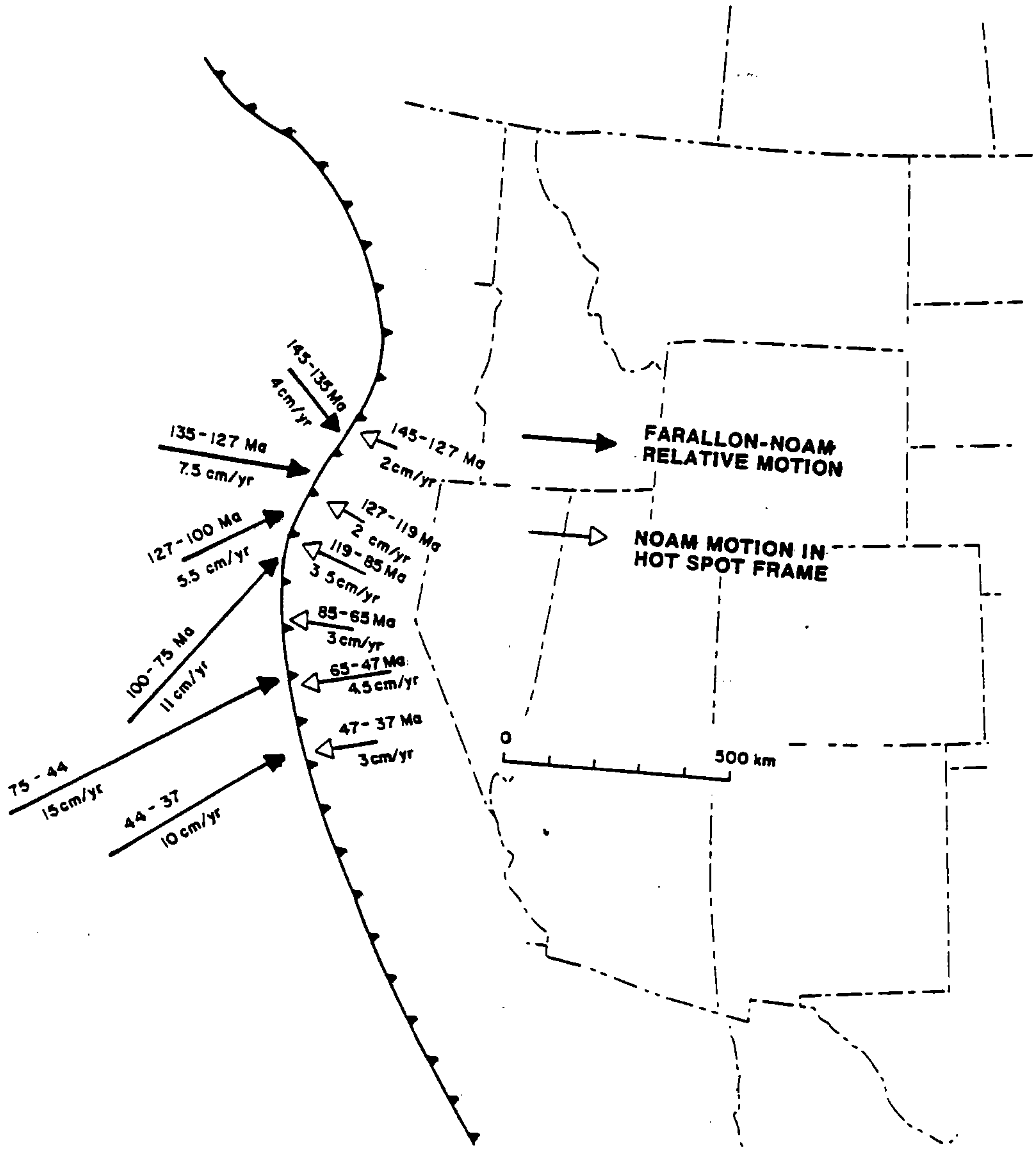


Fig. 1. Schematic representation of North American and Farallon plate motions from 145 to 37 Ma. The inferred position of the subduction zone along the western United States is indicated by the thrust symbol. Open arrows show orientation and velocity of North America in the absolute motion or hot spot reference frame. Closed arrows show relative convergence velocities and orientations of the North American and Farallon plates. Arrows are scaled to the velocities of absolute and relative motions, respectively. Plot derived from plate reconstruction parameters of Engebretson (1972).

gence velocity between the Farallon plate and western United States reached a maximum of 150 km Ma^{-1} and the orientation of convergence was perpendicular to the subduction zone. This combination of high convergence rate and rapid overriding of the trench by North America in the absolute motion frame should result in a shallower angle of subduction. With shallow subduction there should be a concomitant shift of the volcanoplutonic arc away from the trench or, alternatively, cessation of magmatism (Cross & Pilger, 1982).

Cross & Pilger (1978b) and Livaccari, Burke & Sengor (1981) speculated that, during approximately the same time period, an aseismic ridge on the

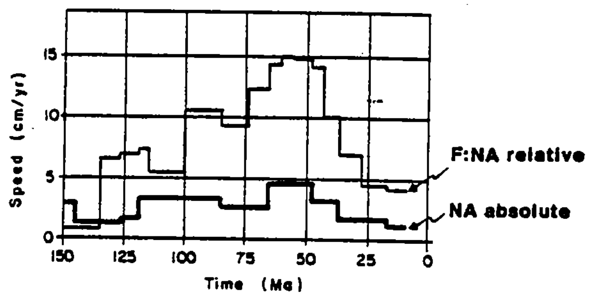


Fig. 2. Velocities of North American and Farallon plate motions from 150 to 10 Ma. Heavy line represents the velocity of North America in the hot spot reference frame. Light line represents the relative convergence velocity between the Farallon and North American plates in a direction perpendicular to an assumed plate boundary trending N40°W. Figure redrawn from Engebretson *et al.* (1984).

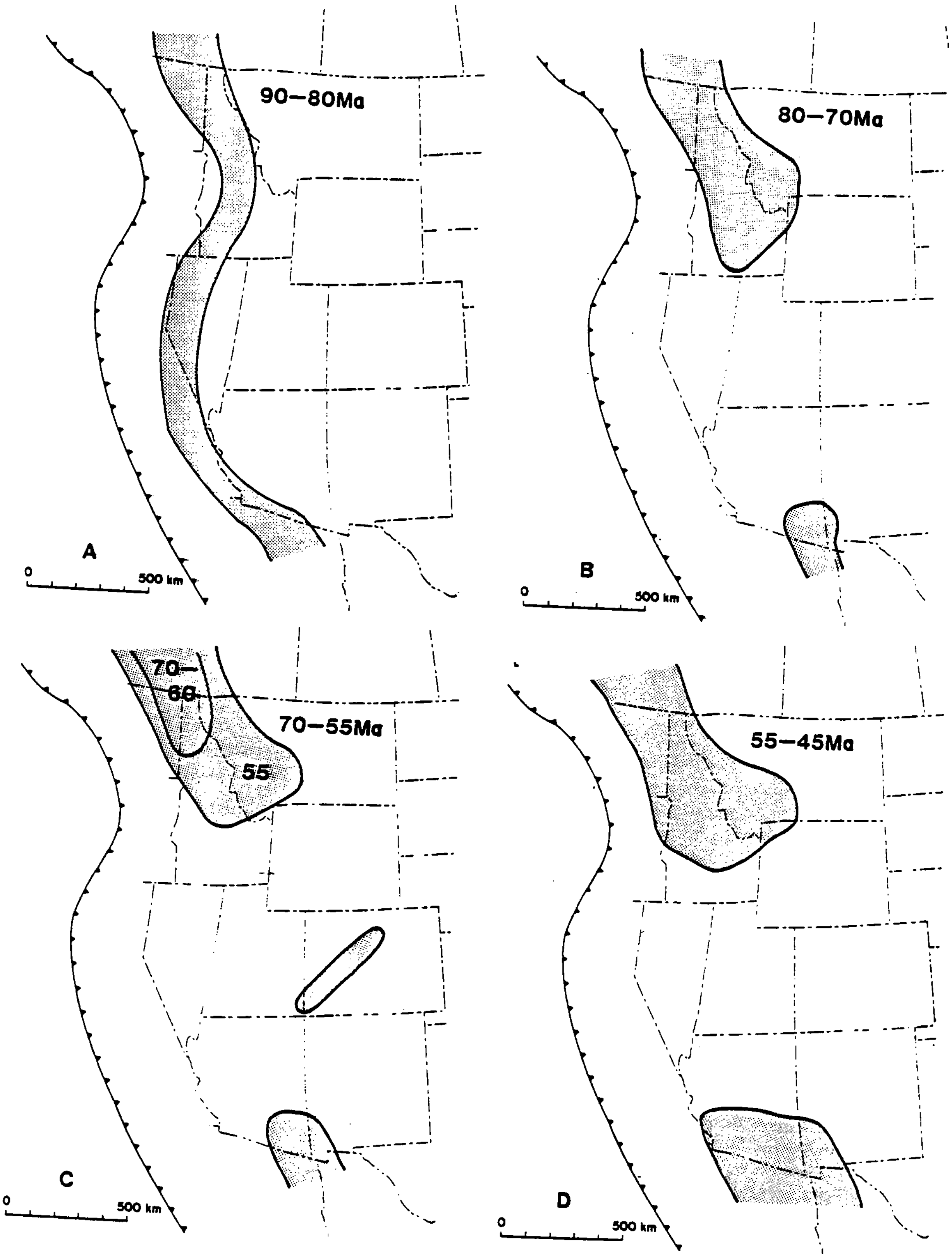
Farallon plate was subducted beneath the central part of western United States. They argued that the increased crustal thickness beneath the aseismic ridge and consequent decreased average lithospheric density made that segment of the Farallon plate more buoyant and either contributed to or caused the development of an anomalously low-angle subduction segment. Subsequent refinements in techniques and approaches of plate reconstructions have provided the means of testing these speculations. Henderson *et al.* (1984) reconstructed the positions of aseismic ridges on the Farallon plate by assuming a mirror imagery of aseismic ridge formation on the Pacific (observable) and Farallon (subducted and not observable) plates. Their reconstructions show that, if it existed, an aseismic ridge intersected the trench near Cape Mendocino, California by 65 Ma. During the following 15 Myr the intersection of the postulated aseismic ridge and the North American plate margin moved rapidly to the south. As a consequence of the relative motions between North America and the Farallon plate, the subducted ridge was beneath portions of the central part of western United States from at least 65–55 Ma. Owing to uncertainties in the plate reconstructions, the location of the thermal anomaly that created the ridge, and the maximum age of the ridge, a 10 Myr error in timing of intersection is permitted. From their discussion of these uncertainties, it is evident that any errors are most likely to shift the time of initial intersection to an earlier date. Subduction of an aseismic ridge should decrease the angle of subduction, extinguish the arc above the subducted ridge, and increase the effectiveness of coupling and stress transmission between the subducting and overriding plates. (Shallow-angle subduction from any other cause would produce the same effects.)

At about 45 Ma, the velocity of the North American plate decreased to about 30 km Ma^{-1} and, concurrently, the relative convergence velocity between North America and the Farallon plate dropped to about 100 km Ma^{-1}. Relative convergence velocity continued to decrease rapidly to a minimum of about 50 km Ma^{-1} by 25 Ma. The anticipated result of these plate interactions is renewed normal subduction and gradual resumption of a coastal volcanoplutonic arc. Slow absolute motion of the North American plate toward the trench combined with slow convergence should create an extensional stress regime in the lithosphere of the upper plate. This regime would be expressed by regional extension in the weak, ductile portions of the crust previously heated by widespread magmatism.

PLATE INTERACTIONS INFERRED FROM HISTORY OF MAGMATISM

The Late Mesozoic and Cenozoic history of igneous activity in the western United States has been described and summarized by Christiansen & Lipman (1972), Lipman, Prostka & Christiansen (1972), Armstrong & Suppe (1973), Snyder, Dickinson & Silberman (1976), Cross & Pilger (1978a) and Lipman (1980). Subsequently published data confirm the space–time patterns of magmatism described in those reports.

These and other studies, based primarily on compilations of isotopic age determinations and compositions of igneous rocks, invoke certain assumptions about genetic relationships between the nature of plate interactions and the generation and composition of magma. From these assumptions and the observed space–time pattern of magmatism, there has emerged a coherent picture of the evolution of plate boundaries and the history and nature of plate interactions along the western United States. The assumptions and the inferred histories of plate interactions have been corroborated in at least two important ways. First, the observed distribution of Neogene, subduction-related igneous rocks is coincident with that predicted from independently derived plate reconstructions (e.g. Snyder *et al.*, 1976; Cross & Pilger 1978a). Second, the inferred plate motion history accounts for a variety of tectonic features and events not directly related to


90–80Ma
A
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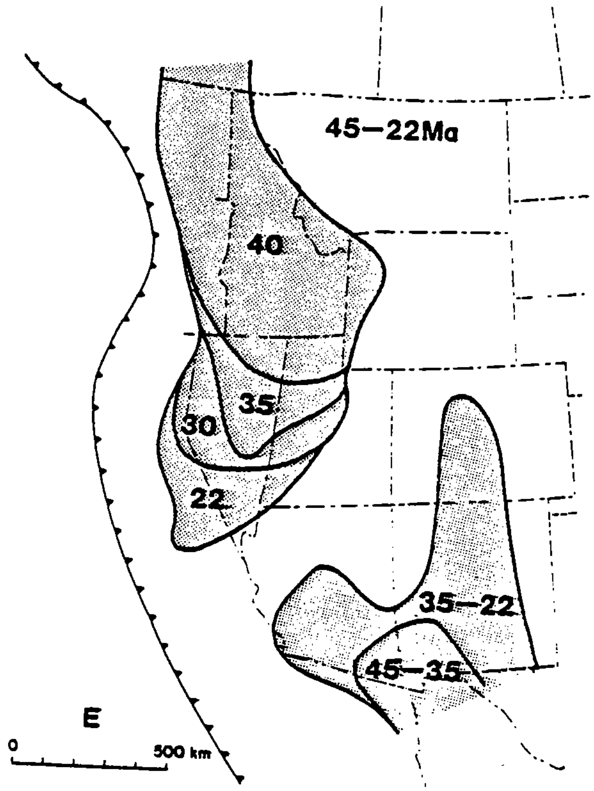


Fig. 3. Summary of history of magmatism in the western United States. Stipple patterns show positions of volcanoplutonic arc as inferred from isotopic ages of calc-alkaline igneous rocks of silicic and intermediate composition. Ages in million years (Ma) indicated on each figure show geographic limits of the inferred arc for the respective times. Palinspastic map of western United States with Late Cenozoic extension and strike-slip faulting removed is from Hamilton (1978). The inferred position of the subduction zone is indicated as in Fig. 1. (A) From Late Jurassic to about 80 Ma, the Sierra Nevadan volcanoplutonic arc was active along the western margin of the North American plate, indicating a period of normal subduction. (B) During the 80–70 Ma period, a broad swath across the central portion of western United States was devoid of magmatism. Magmatism within the Andean-type volcanoplutonic arc continued to the north and south of this magmatic gap. The magmatic gap is inferred to represent a period of shallow subduction, with normal subduction continuing to the north and south of the gap. (C) The magmatic gap in the central part of western United States continued from 70 to 55 Ma, with the exception of minor volcanism along the Colorado Mineral Belt. A genetic association of the Colorado magmatic episode with subduction is not established. (D) The magmatic gap continued in essentially the same position until about 45 Ma. (E) Beginning at 45 Ma, magmatic centres migrated from the north and south into the region previously occupied by the magmatic gap. Thus, the magmatic gap was progressively diminished in size and by 22 Ma magmatism was renewed along essentially the entire margin of western United States. This is inferred to reflect renewed normal subduction. Magmatism in Colorado and New Mexico from 35 to 22 Ma may reflect either regional extension or partial melting of the former shallowly subducted plate after it detached from the upper plate and sank under gravitational force.

magmatism (e.g. Armstrong, 1974; Cross & Pilger, 1978a).

Because the history of igneous activity and its implications for the history and nature of plate interactions are treated in detail in other reports, only a synopsis is given here. Throughout the Jurassic and most of the Cretaceous, igneous activity was confined to a narrow belt along the western margin of the North American plate. Here was developed an Andean-type volcanoplutonic arc above a steeply dipping, eastward verging subduction zone (Fig. 3A; Hamilton, 1969).

At 80 Ma, magmatism waned dramatically in the Sierra Nevadan province of California, Nevada and southern Oregon, and ceased by 70 Ma (Evernden & Kistler, 1970; Chen & Moore, 1982). With the extinction of the Sierra Nevadan arc in this region, a broad swath devoid of magmatism occupied the central part of western United States from about 80 to 72 Ma (Fig. 3B). To the north and south of this magmatic gap, coastal volcanoplutonic arcs extended from Canada into northern Washington and Idaho, and from Mexico into southern Arizona and New Mexico. Between 80 and 70 Ma, the volcanoplutonic arc in the Pacific Northwest expanded and migrated south-eastward into south-western Montana. Active magmatism in southern Idaho and Montana ceased at 70 Ma. A similar, but less pronounced, expansion and eastward shift occurred along the southern volcanoplutonic arc in Mexico and southern Arizona. Coney (1972) recognized that these major shifts in the loci of igneous activity were coeval with equally fundamental changes in tectonism and attributed these changes to reoriented and accelerated motion of the North American plate.

Minor plutonism and associated volcanism occurred along the Colorado mineral belt from 72 to 55 Ma (Fig. 3C). However, a genetic relationship between these small magmatic centres and subduction is not established. With the exception of this minor episode of igneous activity in Colorado, the magmatic

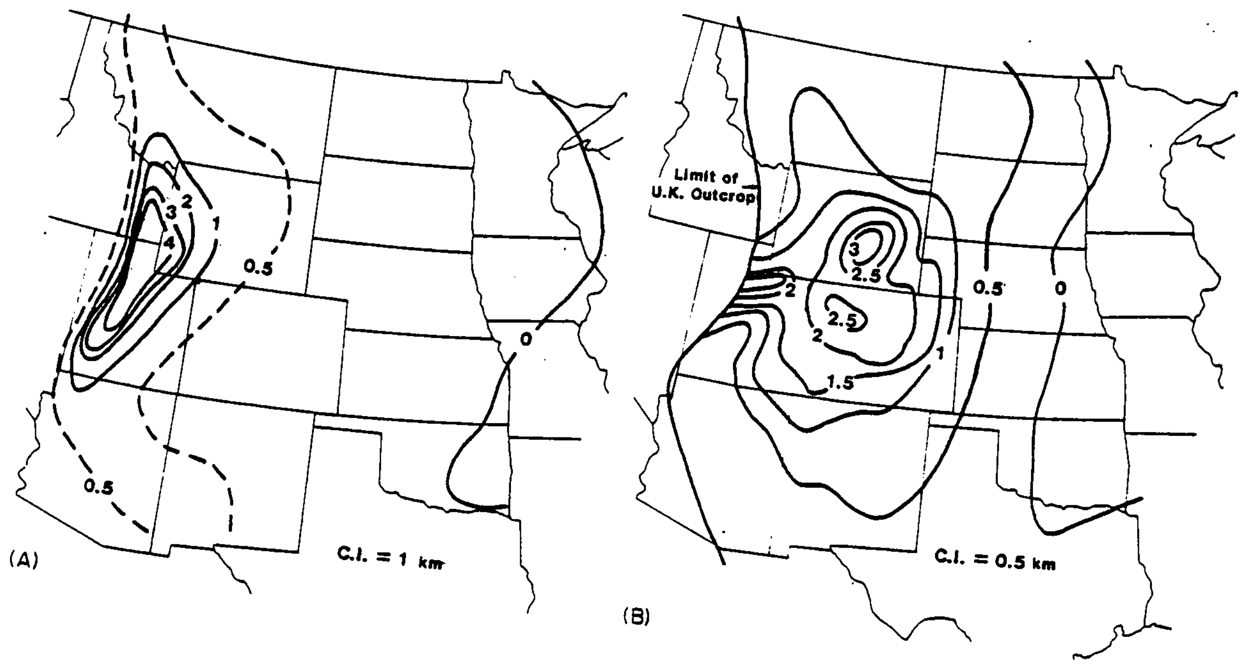


Fig. 4. Two spatially disjunct modes of Cretaceous subsidence are recognized in the foreland of the western United States. (A) One mode was confined to a north–south trending, relatively narrow, asymmetric structural trough. This was colinear with and immediately adjacent to the Sevier fold and thrust belt. Subsidence of the foreland basin (strict sense) is attributed to loading by thrust sheets and derived sediments as discussed in the text. Isopachs show restored thickness of Upper Albian through Santonian strata. (B) The other mode of subsidence occurred during the Campanian and Maastrichtian (84–66 Ma). It was confined to a broad region centred about Wyoming and Colorado. Supracrustal loading by thrust sheets and sediment is insufficient to cause the observed subsidence. Instead, subsidence is attributed to subcrustal loading and cooling by a shallowly subducted oceanic plate. The geographic position and timing of this mode of subsidence in the foreland is coincident with the location and timing of shallow subduction inferred from the history of igneous activity and the development of the magmatic gap. Isopachs show restored thickness of Campanian and Maastrichtian strata. Figure modified from Cross & Pilger (1978b).

gap in the central part of western United States continued to about 45 Ma (Fig. 3D). During the 60–45 Ma period, the northern limit of the magmatic gap is clearly defined by centres of volcanism and plutonism in southern Idaho and south-western Montana. The southern limit of the magmatic gap was essentially stationary from 80 to 45 Ma as defined by continuous igneous activity in southern Arizona and south-western New Mexico. Cessation of magmatism in the Sierra Nevadan arc and development of the magmatic gap has been interpreted in many reports as a consequence of low-angle subduction (e.g. Hyndman, 1972; Lipman *et al.*, 1972; Coney & Reynolds, 1977; Snyder *et al.*, 1976; Cross & Pilger, 1978a; Keith, 1978).

After 45 Ma, magmatism migrated from the north and south to fill the previously amagmatic area (Fig. 3E). Southward expansion of the arc from southern Idaho was rapid, such that by 35 Ma a volcanoplutonic arc extended continuously from the Pacific Northwest to southern Nevada. Northward expansion of the arc from southern Arizona was slower and less pronounced. By 25 Ma, a volcanoplutonic arc again occupied the western margin of the United States, except for a narrow swath in southernmost Nevada and adjacent regions to the east and west. During the same period (45–25 Ma), magmatism was renewed in the southern Rocky Mountains in Colorado and New Mexico. This episode of igneous activity was geographically and volumetrically restricted from 45 to 35 Ma, but expanded rapidly between 35 and 25 Ma. Expansion of magmatism into the central part of western United States has been interpreted in numerous reports (for example, references cited in the preceding paragraph) as a consequence of renewed normal subduction, or flexure and steepening of the previously shallowly subducted lithosphere. Cross & Pilger (1978a) attributed this episode of igneous activity in the southern Rocky Mountains to decoupling of the shallowly subducted Farallon plate segment from the overlying North American plate, and consequent gravitational sinking, heating and magma generation.

PLATE INTERACTIONS INFERRED FROM SUBSIDENCE IN THE FORELAND

Cross & Pilger (1978b) recognized two spatially disjunct modes of subsidence in the foreland basin of the western United States. One mode, considered the foreland basin in the strict sense, was confined to a north–south trending, relatively narrow, asymmetric structural trough colinear with and immediately adjacent to the Sevier fold and thrust belt (Fig. 4A). Cretaceous to Early Cenozoic subsidence of the Wyoming and Utah sector of the foreland basin was described by Jordan (1981). During evolution of the basin, the depositional and structural axis migrated eastwards in response to the eastward advance of thrust sheets and sediment loads. Through numerical modelling, Jordan demonstrated that flexure of lithosphere in response to loading by thrust sheets and derived sediments is sufficient to explain the observed subsidence history and the asymmetry of this sector of the foreland basin.

The other mode of subsidence occurred during the Campanian and Maastrichtian and was confined to a broad region, well to the east of the foreland basin proper, in southern Wyoming and western Colorado (Fig. 4B). Isopachs of Campanian and Maastrichtian strata are sub-circular in map pattern and depict centres of maximum thicknesses (3 km) displaced approximately 300 km to the east of the foreland basin structural axis. Subsidence in the Colorado–Wyoming locus was coeval with subsidence of the foreland basin (strict sense) to the north and south, and Jordan's (1981) study showed that supracrustal loading by thrust sheets and sediment is insufficient to cause the subsidence. (The Campanian and Maastrichtian stages span approximately 24 Myr, a time span greater than is desirable for directly comparing the history of igneous activity, particularly the development of the magmatic gap, with the history of subsidence in the foreland. However, complexity in stratigraphic nomenclature and correlation, and insufficient numbers of isotopic age dates of stratigraphic units over the region have precluded making isopach maps of stratigraphic units representing shorter time intervals. In general, lower Campanian units are thinner than upper Campanian units. Therefore, it is likely that this second mode of subsidence in the foreland developed late in the Campanian and continued through the Maastrichtian.)

Cross & Pilger (1978b) discussed the temporal and spatial correspondence between the Colorado–Wyoming locus of subsidence and the postulated occurrence of a shallowly subducted Farallon plate segment. The inferred geometry of this episode of shallow subduction placed cold oceanic lithosphere in close contact with overlying continental lithosphere, thus displacing hot, less dense asthenosphere. As a consequence of the changed geometry and vertical density structure, Cross & Pilger proposed that two mechanisms, sublithospheric loading and sublithospheric cooling, were sufficient to cause the observed subsidence. Their calculations, assuming perfectly elastic Airy compensation, indicated isostatic (rapid) subsidence of 3 km due to sublithospheric loading. Subsidence due to cooling of the base of the continental lithosphere (slow) was an additional 2 km.

Bird (1984) conducted a rigorous, quantitative test of these proposed mechanisms by Fourier-transform solution of the plate bending equation and by finite-difference thermal modeling. His calculations confirmed that the additional weight of the subducted plate would have depressed the region by at least the observed amount of subsidence. The magnitude of subsidence calculated from his model is as much as 1 km greater than that recorded by stratal thicknesses, and the area of maximum subsidence predicted by the model is broader than that observed. (Note, however, that the stratal thicknesses reported by Cross & Pilger (1978b) were uncorrected for compaction. Conservatively assuming an average loss of 40% porosity during burial of the predominantly shale and mudstone sedimentary section, the cumulative sedimentary thickness of a 3 km column of strata would have been 4.2 km.) Nonetheless, a remarkable correspondence exists between the modelled and the observed subsidence, given that the region of greatest modelled subsidence occurs west of the Sevier fold and thrust belt, an area of initially high elevation that lacks a sedimentary record. In particular, Bird (1984) noted that the 0 and 1 km subsidence contours of his model are essentially coincident with the observed 0 and 1 km isopach contours, and the patterns of the modelled and observed contours are essentially conformable. Bird also noted that subcrustal cooling would augment the subsidence due to subcrustal loading, although its effect would have been temporary and the additional amount of subsidence was not reported.

SEVIER AND LARAMIDE OROGENIES

Background

Late Mesozoic and Early Cenozoic deformation in the Rocky Mountains region traditionally has been categorized as two temporally distinct compressional events termed the Sevier and Laramide orogenies. The former traditionally is regarded as a dominantly intra-Cretaceous event characterized by thin-skinned thrusts and folds of décollement style, which usually, but not exclusively, did not involve basement. Deformation along the Sevier foreland fold and thrust belt was confined to strata of the westward thickening Palaeozoic and Mesozoic sedimentary prism. The most eastern deformational limits of the Sevier fold and thrust belt correspond to the approximate positions of the Palaeozoic craton-margin hingelines. By contrast, the Laramide orogeny traditionally is regarded as a latest Cretaceous to Early Cenozoic event that affected the craton. Laramide structures are characterized by basement-cored uplifts and asymmetric anticlines, typically bounded by high-angle reverse and thrust faults.

Armstrong (1974) and Burchfiel & Davis (1975), among others, drew attention to the lack of a clear temporal distinction between the two events, if they are regarded from the perspective of differing kinematics and styles of strain. They noted that basement-involved deformation of Laramide age was confined to the southern and central Rocky Mountains of New Mexico, Colorado, Wyoming and southern Montana. In this sector of the Rocky Mountains, deformation along the Sevier fold and thrust belt ceased shortly before initiation of Laramide basement-involved deformation. To the north (in Montana, Idaho and Canada) and to the south (in west Texas, New Mexico, Arizona and Mexico), thin-skinned deformation of the fold and thrust belt was continuous from latest Jurassic(?) to Eocene. In summary, basement-involved deformation was latitudinally restricted to the central and southern Rocky Mountains and was contemporaneous with continued thin-skinned deformation to the north and south. Although the terms Laramide and Sevier are traditionally used as temporal distinctions of deformational events throughout the Rocky Mountains (and, the terms often are used as temporal designations for events in other parts of the world as well), it is clear that such usage should be abandoned. Instead, the terms should be restricted to spatially and temporally restricted orogenic events characterized by differing styles of deformation in the Rocky Mountains.

Armstrong (1974) noted the coincidence among three major events: the cessation of Sevier style deformation in Utah and Wyoming; the initiation of Laramide style deformation in the central and southern Rocky Mountains; and, the history of igneous activity in the western United States. He attributed the abrupt change from thin-skinned to basement-involved deformation and the attendant eastward displacement of deformation to a ductility contrast within the continental lithosphere. Sevier style deformation was coeval and latitudinally conterminous with Andean type volcanism and plutonism. The responses to compressive stresses were predominantly ductile shortening within the crust heated by magmatism, and detachment and internal imbrication of the overlying, westward thickening sedimentary prism. Cessation of igneous activity in the Sierra Nevadan arc and formation of the magmatic gap at 80 Ma caused the previously high heat flow of the arc and back-arc region to dissipate. Consequently, the lithosphere in the vicinity of the magmatic gap cooled, thickened and became resistant to deformation as a consequence of its increasing strength. Thereafter, compressive stresses were relieved by buckling and shear of the more brittle upper crust. By contrast, the lithosphere to the north and south of the magmatic gap, opposite the volcanoplutonic arcs, remained hot and more ductile. Armstrong's explanation adequately accounts for the differences in styles of deformation and their regional distributions. However, it neither accounts for the abrupt eastward displacement of deformation, nor for the curvature and specific loci of Laramide structures.

Laramide structures of the southern Rocky Mountains are oriented approximately north–south, coincident with the eastern margin of the Colorado Plateau (Figs 5 and 9D, E). These occupy a band extending about 200 km east from and parallel to the physiographic limit of the Plateau. As Laramide structures are traced into the central Rocky Mountains, structural trends rotate anticlockwise first to NW–SE and then to east–west. Orientations of these Laramide structures conform closely to the northern and northeastern margins of the Plateau. With the exception of the Uinta Mountains, Laramide structures of the central Rocky Mountains also form a band about 200 km in breadth, but which is displaced about

300 km north and NE from the physiographic margin of the Plateau.

Hamilton (1978, 1981) revived earlier speculations about a possible genetic relationship between Laramide structures and their spatial association with the margins of the Colorado Plateau. He explicitly noted the parallelism between orientations of the margins of the Colorado Plateau and orientations of Laramide structures. He reasoned that the relatively undeformed Plateau behaved as a mechanically rigid body, and that Laramide deformation was an expression of compression between a Plateau microplate and the larger plate of the continental interior. He described kinematically the relative motion between the two plates as a rotation 2°–4° about an Euler pole located to the SE in the Texas Panhandle. Finite rotation of two plates about a geometric pole only may describe the sum of their relative motions over a discrete time interval. However, as recognized by Hamilton, the geometric pole need not have remained fixed and the actual relative plate motions through time (as potentially described by instantaneous rotations about moving poles) may have differed in important ways from the sum of the finite rotation.

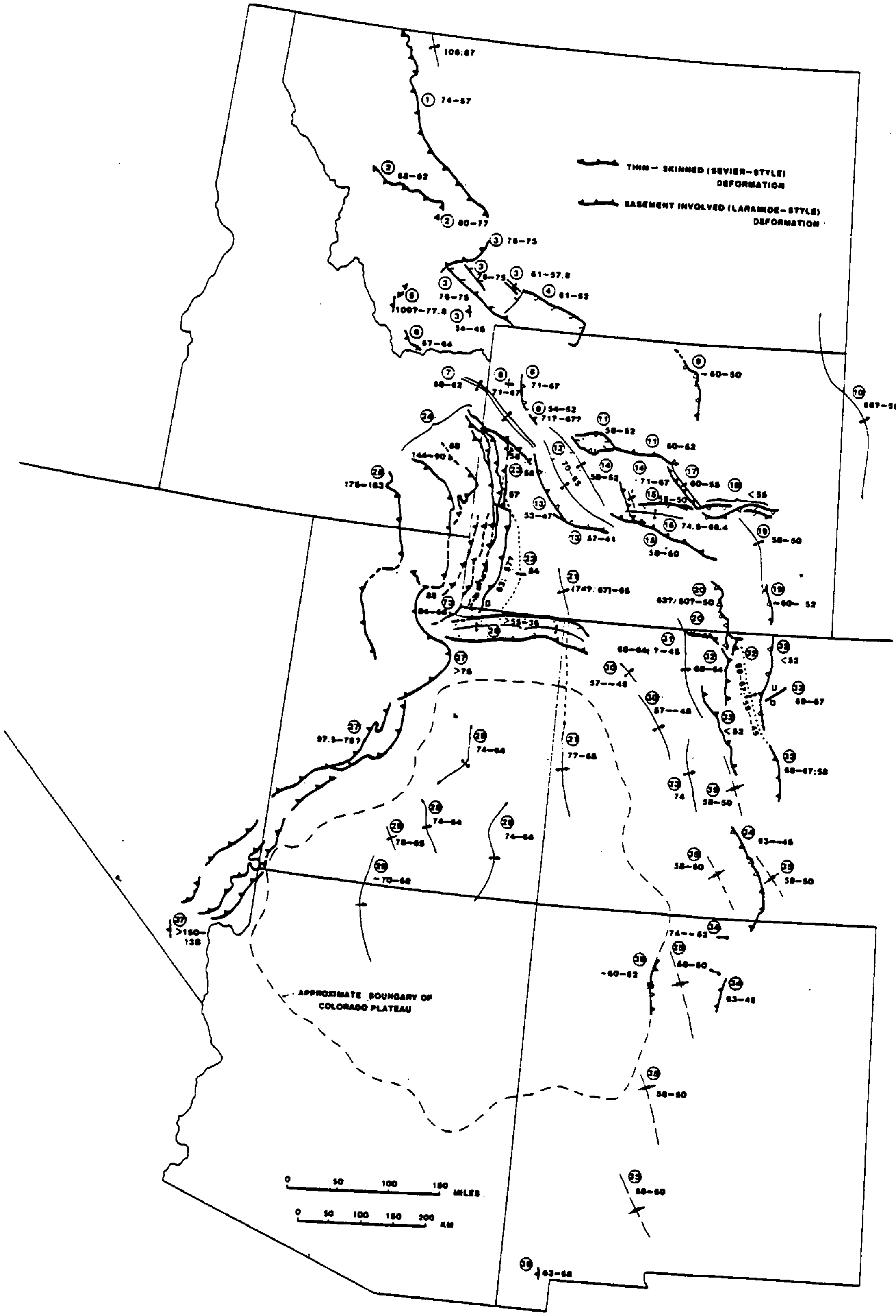
Gries (1983) criticized the kinematic description of Hamilton on the basis that it failed to account for perceived differences in the amount and timing of crustal shortening along east–west trending (greater and later, respectively) and north–south trending (lesser and earlier, respectively) Laramide structures. Instead, she proposed an alternative kinematic description of relative motion between the Colorado Plateau microplate and the continental interior. Gries suggested that relative motion during Late Cretaceous through early Palaeocene was east–west, whereas, during the late Palaeocene through late Eocene, it was north–south.

Regardless of the eventual resolution of finite or instantaneous rotations describing the motions of the Plateau relative to the continental interior, the more basic questions of causation and mechanism remain unaddressed. If, as seems probable, Laramide structures are geometrically and kinematically related to the Colorado Plateau, what was responsible for the origin and maintenance of the Plateau as a relatively rigid tectonic block? What caused the abrupt eastward displacement and change in style of deformation from the Sevier to the Laramide belts in the central portion of western United States? What controlled the development of Laramide structures along a 200 km broad zone parallel with the northern and eastern margins of the Plateau? What other Late Mesozoic and Early Cenozoic tectonic events and elements may be related genetically to the change from thin-skinned to basement-involved deformation?

Duration and geographical limits of Sevier deformation

Resolution of these questions ultimately requires specific knowledge of the nature, position and timing of deformation. Consequently, a set of geological data that places temporal and spatial limits on deformation in the Rocky Mountains was compiled from pertinent literature. From among the scores of major structural elements that have been described in a very large literature, only a few dozen were selected. Only those structural elements for which the initiation, duration, and/or cessation of deformation are well defined within narrow time limits by cross-cutting relations and palaeontologic or isotopic age determinations are included in the compilation. All stratigraphic, biostratigraphic, and isotopic ages were converted to the 1983 Decade of North American Geology geological time-scale (Palmer, 1983). In all instances, primary literature was examined. Results of this compilation are presented in the map of Fig. 5. Comparison of this map with other reviews of more generalized or more limited temporal and geographic scope will reveal occasional discrepancies in both the structural elements selected as defining times of deformation and the ages of deformation assigned to specific structures. These discrepancies are attributed to repeated usage of derivative, rather than primary, literature and, in a few instances, to less than critical analysis of the data from which interpretations were derived.

The limits of initiation and cessation of Sevier deformation in portions of the Rocky Mountains remain elusive, particularly in the critical region adjacent to the belt of Laramide structures. Ages of thrusting are best documented along the Overthrust Belt of Wyoming, south-eastern Idaho and northern Utah (see review and analysis of Wiltschko & Dorr, 1983). There is good evidence for thrusting of the Crawford and Meade allochthons at 88 Ma. Earlier movement along the Paris and Willard thrusts, beginning at 144 Ma and continuing episodically until 90 Ma, is indicated by conglomeratic strata interpreted as synorogenic clastic wedges. Stratigraphic relations in central Utah (Fouch *et al.*, 1983; Lawton & Mayer, 1982) provide substantial evidence for initiation of thrusting as early as Cenomanian (97·5 Ma). Armstrong (1968) reviewed more specula-


THIN-SKINNED (SEVIER-STYLE) DEFORMATION
BASEMENT INVOLVED (LARAMIDE-STYLE) DEFORMATION
APPROXIMATE BOUNDARY OF COLORADO PLATEAU
0 50 100 150 MILES
0 50 100 150 200 KM

tive evidence that suggested that deformation along the central and southern Utah portion of the Sevier belt began in the Early Cretaceous. Even earlier Middle Jurassic (175–163 Ma) dislocation along the Manning Canyon Shale detachment in northern Utah and southern Idaho was inferred by Allmendinger & Jordan (1981). The earliest siliclastic sediments definitely derived from the west occur within the Stump Formation (Oxfordian; 156–163 Ma) in western Wyoming, SE Idaho and northern Utah (Brenner 1983). The earliest clastic wedge in the same area occurs within the Morrison Formation (Ephriam Conglomerate) of Oxfordian to Kimmeridgian age (Brenner, 1983). Although these strata were derived from the west, their association with deformation of the Sevier fold and thrust belt has not been established.

In Montana, the earliest well documented occurrence of thin-skinned deformation is thrusting of the Sapphire plate between 88 and 82 Ma (Ruppel, 1963; Ruppel *et al.*, 1981). Earlier deformation beginning by 100 Ma in SW Montana is suggested by ages of the Beaverhead Formation interpreted as a synorogenic deposit (Ryder & Scholten, 1973). More recent palynologic studies of the Beaverhead Formation by Nichols, Perry & Haley (1985) suggest that deformation did not begin until early Campanian (84 Ma). Lorenz (1982) inferred from an episode of rapid subsidence along the Sweetgrass Arch that initiation of deformation in NW Montana began about 106 Ma.

Along the southernmost extent of the Sevier belt in Nevada, ages of thrusts (Davis, 1973) and of inferred syntectonic conglomerates (Carr, 1980) indicate that deformation began in latest Jurassic (≥ 150 Ma) and continued episodically to 138 Ma.

In summary, structural and stratigraphic relations document that thin-skinned deformation was occurring throughout the Sevier fold and thrust belt by about 90 Ma. Although evidence for earlier initiation

Fig. 5. Palaeotectonic map showing positions and age limits of deformational events that are well constrained by available evidence. Sources of information for localities indicated on the map are as follows. **Locality 1** (Montana Disturbed Belt: Glacier National park: NW Montana): Hoffman, Hower & Aronson (1976); Lorenz (1982); Mudge (1972a,b, 1980, 1982); Robinson, Klepper & Obradovich (1968); Schmidt (1978). **Locality 2** (Saphire plate): Ruppel (1963); Ruppel *et al.* (1981). **Locality 3** (SW Montana): Schmidt & Garihan (1983). **Locality 4** (Beartooth Mountains): Foose, Wise & Garabarini (1962). **Locality 5** (Grasshopper and Medicine Lodge plates): Hammons (1981); Perry, Ryder & Maughan (1981); Ruppel *et al.* (1981); Thomas (1981). **Locality 6** (Tendoy thrust): Hammons (1981); Perry & Sando (1982); Perry *et al.* (1981); Ryder & Scholten (1973). **Locality 7** (Targhee Uplift; Ancestral Teton-Gros Ventre Highland): Dorr, Spearing & Steidtmann (1977); Love (1973); Love, Leopold & Love (1978); Wiltschko & Dorr (1983). **Locality 8** (SW flank, Washakie Range): Keefer (1965a, b); Love (1973); Winterfeld & Conard (1983). **Locality 9** (east flank, Bighorn Mountains): Curray (1971). **Locality 10** (Black Hills): Lisenbee (1978); Love (1960). **Locality 11** (southern Bighorn Mountains; southern Owl Creek Mountains; northern Wind River Basin; Casper Arch): Gries (1983); Keefer (1965a, b); Keefer & Love (1963); Love (1978). **Locality 12** (Wind River Range; Wind River Basin): Berg (1963); Gries (1983); Keefer (1965a,b, 1970); Keefer & Love (1963); Love (1960, 1970). **Locality 13** (south flank, Wind River Range): Gries (1983); Love (1970); Steidtmann, McGee & Middleton (1983). **Locality 14** (west flank, Granite Mountains; Emigrant Trail thrust): Keefer (1965a, b); Love (1970, 1971); Reynolds (1978). **Locality 15** (north and south flanks, Granite Mountains; north flank, Great Divide Basin): Gries (1983); Love (1970, 1971); Reynolds (1976, 1978). **Locality 16** (Sweetwater Arch): Love (1960, 1970); Reynolds (1971). **Locality 17** (Casper Arch): Keefer (1965a, b); Keefer & Love (1963); Love (1978). **Locality 18** (north flank, Laramie Range; Casper Mountain): Keefer (1965a, b, 1970). **Locality 19** (NW and east flanks, Laramie Range): Blackstone (1975); Tweto (1975). **Locality 20** (Medicine Bow Mountains): Blackstone (1975); Knight (1953); Tweto (1975). **Locality 21** (Ancestral Rock Springs Uplift; Douglas Creek Arch): Ritzma (1955); Roehler (1961). **Locality 22** (Moxa Arch): Thomaidis (1973); Wach (1977); Wiltschko and Dorr (1983). **Locality 23** (Cache thrust): Kopania (1983); Wiltschko & Dorr (1983). **Locality 24** (northern Sevier fold and thrust belt): Wiltschko & Dorr (1983, and references therein). **Locality 25** (Manning Canyon detachment): Allmendinger & Jordan (1981). **Locality 26** (Uinta Mountains): Hansen (1984); Ritzma (1955, 1971); Standlee (1982). **Locality 27** (central Sevier fold and thrust belt): Armstrong (1968); Standlee (1982); Lawton (1983); Lawton & Mayer (1982). **Locality 28** (San Rafael Swell; Circle Creek Uplift; Monument Upwarp): Lawton (1983). **Locality 29** (East Kaibab and Dutton Mountain monoclines): Bowers (1972); Kelley (1955); **Locality 30** (Axial Arch; White River Plateau): Tweto (1975). **Locality 31** (Park Range): Chapin & Cather (1981); Tweto (1975). **Locality 32** (Front Range): Blackstone (1975); Chapin & Cather (1981); Tweto (1975); Weimer & Davis (1977). **Locality 33** (Sawatch Uplift): Tweto (1975). **Locality 34** (Sangre de Cristo and San Luis Highlands): Chapin & Cather (1981); Dickinson, Leopold & Marvin (1968); Tweto (1975). **Locality 35** (eastern margin of Colorado Plateau): Chapin & Cather (1981). **Locality 36** (Nacimiento Uplift; east flank, San Juan Basin): Baltz (1967); Kelley (1955); Woodward, Kaufman & Anderson (1972). **Locality 37** (southern Sevier fold and thrust belt): Carr (1980); Davis (1973). **Locality 38** (Little Hatchet Mountains): Loring & Loring (1980).

of Sevier deformation is either geographically restricted or more speculative, available data strongly suggest that deformation was widespread by 100 Ma.

An alternative method of defining initiation and duration of deformation is that of geohistory analysis (Van Hinte, 1978). Lithologies, thicknesses and ages of strata at nine locations along the northern and central portions of the Sevier belt were compiled for geohistory analyses. Results of four of these analyses, representative of the set, are presented in Fig. 6. Interpretation of geohistory curves is ambiguous, because the curves only describe the history of subsidence, not the mechanism which induced the subsidence. The shape of the subsidence history curves is typical for foreland basin subsidence (e.g. Jordan, 1981). Because of this observation and because no evidence exists for an alternative subsidence mechanism, it is assumed that a major increase in subsidence rate (the point of inflection on the subsidence history curve) reflects the initial and subsequent flexure of the foreland basin in front of thrust loads. A limitation of these analyses is that the earliest period(s) of thrusting may not be recorded if the thrust load was more than 200 km west of the selected sites.

A convenient starting point for each curve is Early Jurassic, because Late Triassic and Early Jurassic strata in the Rocky Mountains dominantly comprise aeolianite sediments which are assumed to represent deposition at or near sea-level. At one locality in NW Montana (Fig.6A), the geohistory curve shows one inflection point at 108 Ma and a second inflection point occurs at 89 Ma. The subsidence history curve at one location in SW Montana (Fig. 6B) shows identical positions of inflection points. In the Hoback basin of north-western Wyoming (Fig. 6C), the age of initial loading inferred from the geohistory analysis is 114–107 Ma. The most southerly location along the Sevier belt possessing a stratigraphic record sufficiently complete for geohistory analysis is the Gunnison Plateau region of central Utah (Fig. 6D). The inflection point of this curve is at 97 Ma, although substantial subsidence and sediment accumulation (approximately 5 km) occurred between Early Jurassic and 97 Ma. These analyses support observations from field relationships that deformation throughout the northern and central portion of the Sevier belt was initiated in the late Early Cretaceous.

As previously discussed, cessation of deformation is diachronous along the Sevier belt. In Montana and along the Overthrust Belt of Wyoming, Idaho and northern Utah, thrusting was episodic through the early Eocene. Ages of deformation obtained from most well documented structures indicate that thrusting ceased by 57 Ma, although one thrust in Montana (locality 3, Fig. 5; Schmidt & Garihan, 1983) is bracketed as post-54 and pre-45 Ma. By contrast, structural and stratigraphic relations in the central Sevier belt in Utah indicate that the youngest deformation is late Campanian ($\geq$ 75 Ma). This is supported by the geohistory analysis of strata from central Utah (Fig. 6D). An inflection point at 74 Ma marks a major decrease in subsidence rate from 134 m Ma^{-1} during the 92–74 Ma interval to 25 m Ma^{-1} during the 73–23·7 Ma interval.

Duration and geographical limits of Laramide deformation

The age of cessation of Sevier deformation in central Utah marks the initiation of Laramide deformation (compare Figs 7A and 7B). The earliest Laramide

Fig. 6. Geohistory analyses at four locations along the Sevier fold and thrust belt. Location of each plot is shown on the small inset map. The origin of each curve in the Early Jurassic is assumed to be at sea-level, because Late Triassic and Early Jurassic strata were deposited as widespread, aeolianite sediments. Solid curves represent total subsidence. Dashed curves represent subsidence of the Early Jurassic surface corrected for the incremental load induced by the weight of sediment through time and, thus, the subsidence due to tectonic loading by thrust sheets. Dots on curves mark time/thickness positions at which calculations were made.

Isostatic corrections were based upon Airy's model of isostasy which assumes perfectly elastic response to loading and ignores the flexural strength of the lithosphere (see, e.g. Watts & Steckler, 1979). Corrections for compaction were based upon lithology and follow the exponential porosity functions presented by Sclater & Christie (1980) with the following modifications: grain-supported limestones were treated as sandstones and mud-supported limestones were treated as shaley sand. No corrections for eustatic variation or palaeobathymetry were made.

Sources of data for each curve are: (A) NW Montana (Mudge, 1982, 1972a); (B) SW Montana (Richards, 1957); (C) Hoback Basin, NW Wyoming (Dorr *et al.* 1977; Wanless, Belknap & Foster, 1955); and (D) Gunnison Plateau–Cedar Hills region, central Utah (Jefferson, 1982; Lawton, 1983; Standlee, 1982; Stanley & Collinson, 1979).

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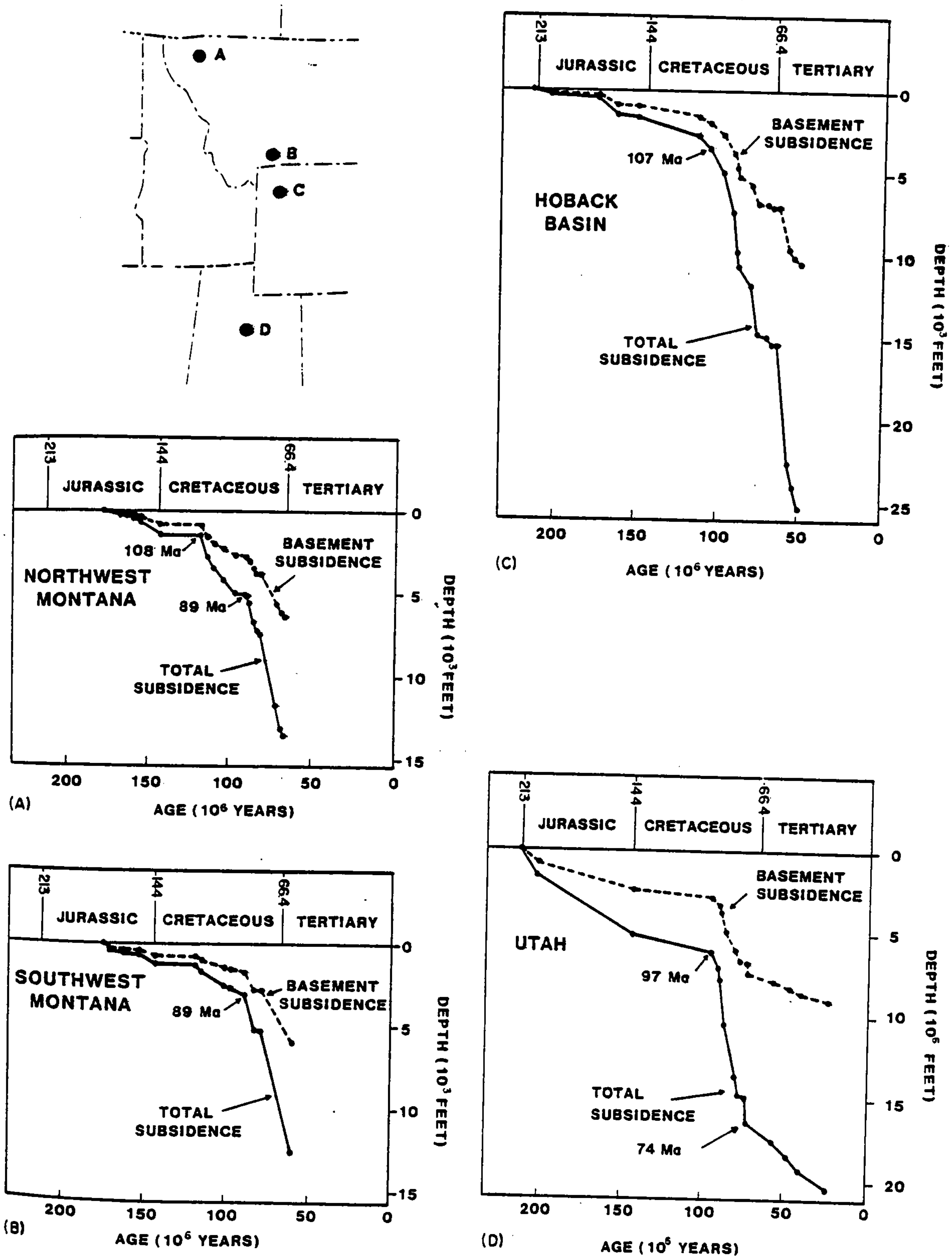

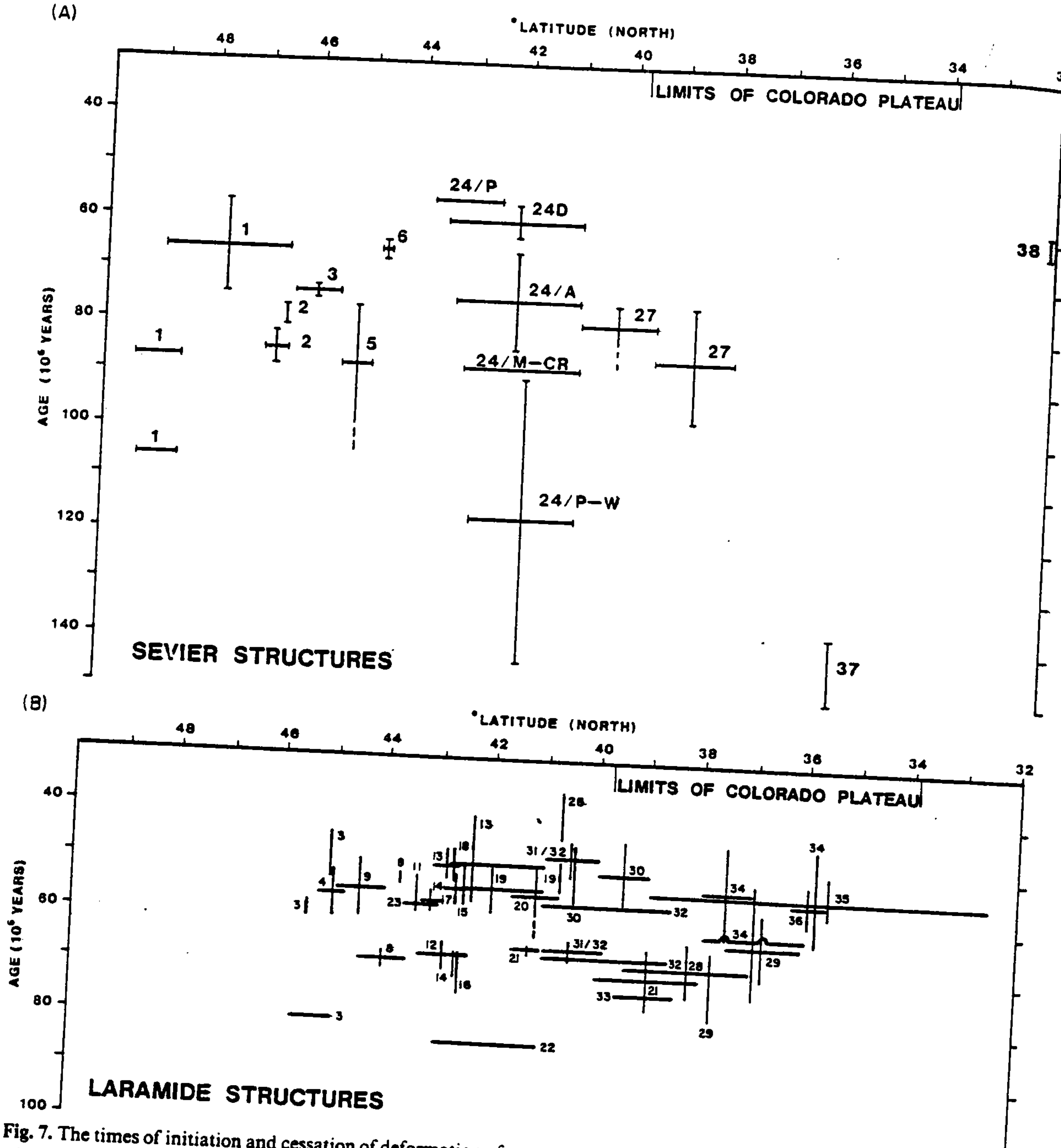

Fig. 7. The times of initiation and cessation of deformation of structures shown on the palaeotectonic map of Fig. 3 are plotted with respect to contemporary latitude. Latitudinal limits of structures are drawn perpendicular to a line of cross-section parallel to and bisecting the trend of the Rocky Mountains (approximately N10°W). The latitudinal limits of the physiographic margins of the Colorado Plateau are identified on each cross-section for geographic reference. (A) Structures associated with Sevier deformation characterized by thin-skinned thrusts and folds of décollement style which typically did not involve basement. (B) Structures associated with Laramide deformation characterized by basement-cored uplifts and asymmetric anticlines, typically bounded by high-angle reverse and thrust faults. The apparent time of hiatus in Laramide deformation is highlighted by the pattern (65–60 Ma).

Laramide structures occupy a space/time frame that differs from that of Sevier structures. Sevier deformation begins earlier than Laramide deformation and initially is continuous along the length of the Rocky Mountains. The earliest Laramide structures developed at or near the beginning of the Maastrichtian (74.5 Ma) at the same time that Sevier deformation in Utah, Wyoming and Idaho ceased. To the north and south of the belt of Laramide style deformation, Sevier style deformation was continuous and coeval with Laramide deformation.

structures developed at or near the beginning of the Maastrichtian (74.5 Ma). Most consist of gentle folds of the Colorado Plateau, the ancestral Rock Springs uplift and Douglas Creek arch, the San Luis Highlands, the Sawatch uplift, and the Sweetwater arch (Fig. 5). An exception to this style of deformation is exhibited by two basement-involved thrusts of earliest Maastrichtian age in SW Montana (Schmidt & Garihan, 1983). Generally, these earliest Laramide structures have a north–south orientation, but some trend east–west and others NW–SE. Several more structures developed later in the Maastrichtian beginning about 69 Ma. These occur along the northern and eastern margins of the Colorado Plateau and again exhibit the same three structural orientations, with the north–south trend being dominant. Two-thirds of these structures are gentle folds and the remainder are basement-involved reverse faults.

If the structures selected in this compilation are representative of the style and timing of deformation of other Laramide structures, the following observations may characterize the early phase of Laramide deformation. Initial deformation was widespread and essentially synchronous throughout the Colorado Plateau and along its margins. Early Laramide deformation is typified by gentle folds and flexures, with minor fault-bounded basement-cored uplifts. By the close of the Cretaceous, the general geographic limits of the Laramide belt were outlined and three structural orientations were defined, the most prominent of which was north–south. Data compiled in this report do not support the speculation of Dickinson & Snyder (1978) that there was a southward migration in the initiation of Laramide deformation. The widespread distribution of early Laramide structures suggests an equally broad distribution and transmission of stress. As discussed subsequently, this widespread distribution stands in marked contrast to the concentration of strain within a relatively narrow curvilinear belt that typifies later Laramide deformation.

There is an apparent hiatus in Laramide deformation during the early Palaeocene. Given the restrictions of the compilation and of the field relations from which temporal limits of deformation were determined, scrutiny of the palaeotectonic map (Fig. 5) reveals that the early phase of Laramide deformation was over by the end of the Cretaceous and that deformation was renewed at 60 Ma. This pause between tectonic episodes is more clearly displayed in the compressed cross-sectional plot of time versus geographic position of deformation (Fig. 7). Because the selection criteria eliminated many structures generally regarded as 'Laramide' in age, this apparent hitus of about 5 Myr duration may be an artefact of a small sample size. Nonetheless, two factors argue for its reality. First, the few structures which extend into the 65–60 Ma period (right side of Fig. 7B) are those with a relatively long range between their established limits of initiation and cessation of deformation; the actual period of deformation may have been shorter. Second, as discussed below, the apparent early Palaeocene hiatus separates structures of different styles and locations of deformation.

Whether or not the apparent tectonic pause is a reality, a second phase of Laramide tectonism began at 60 Ma and continued through the early Eocene. Two phases of Laramide tectonism are distinguished by differing styles and locations of deformation. After the early Palaeocene, Laramide structures are dominated by basement-cored, fault-bounded uplifts. This style of deformation contrasts with the broad, ductile folds characteristic of the early Laramide tectonic episode. Structures formed during the late Laramide phase of deformation are confined to a 200 km-wide belt parallel to the margins of the Colorado Plateau. This also contrasts with the more widespread distribution of early Laramide structures. As noted previously, in the southern Rocky Mountains this belt of late Laramide structures is coincident with the eastern margin of the Colorado Plateau. In the central Rocky Mountains the 200 km-wide belt is displaced about 300 km north and NE from the physiographic margin (but not necessarily the palaeotectonic margin) of the Plateau. In addition, orientations of individual late Laramide structures closely parallel the orientation of adjacent Plateau margins. The parallelism between the physiographic margins of the Colorado Plateau and orientations of structures suggests that the latter were formed by compression at the juncture of a relatively rigid Plateau tectonic block and the continental interior, as suggested most recently by Hamilton (1978, 1981).

Inspection of the palaeotectonic map (Fig. 5) and the compressed cross-section (Fig. 7) reveals that Laramide deformation was renewed at 60 Ma along the eastern and northern margins of the Colorado Plateau. By contrast, the end of Laramide deformation is less well established because few structures possess cross-cutting relations, such as strata deposited upon thrust contacts, that narrowly constrain the cessation of deformation. Many structures have limits of cessation that group around 52–50 Ma; others have limits that group around 45 Ma. Otherwise, deformation apparently was continuous along the curvilinear Laramide belt from 60 Ma to middle Eocene.

Gries (1983) postulated a late Palaeocene lull in Laramide deformation which she regarded as reflecting a change from east–west compression during the Late Cretaceous to early Palaeocene to north–south compression during the Eocene. She proposed that north–south trending structures were formed during the early phase of deformation and east–west trending structures were formed during the late phase of deformation. In this view, the Eocene phase of deformation resulted in greater strain than the earlier phase, and produced significant overlap of basement along east–west oriented structures bounded by reverse faults. The data compiled for this report do not show the late Palaeocene lull in deformation postulated by Gries. Rather, deformation appears continuous from late Palaeocene to the early Eocene. Further, the selected data do not support Gries' conclusion that north–south trending structures were formed first and that east–west trending structures were formed last; structures of both orientations are coeval. However, owing to the nature of this compilation, it is not possible to address whether strain was greater along east–west oriented structures than along north–south oriented structures.

Summary

In summary, the Sevier and Laramide orogenies reflect differing kinematic and dynamic responses to compressive stresses that existed within western North America during the Late Mesozoic and Early Cenozoic. Deformation associated with these events was diachronous along the Rocky Mountains region. The Sevier foreland fold and thrust belt (and its equivalents in Canada and Mexico) occurred opposite an Andean-type volcanoplutonic arc in lithosphere rendered ductile by high heat flow in the arc and back-arc region. Laramide deformation occurred within a region devoid of magmatism and above an inferred shallowly-inclined, subducted oceanic plate. Initiation of Laramide deformation was essentially coeval with cessation of Sevier deformation within the same latitudinal belt; thin-skinned deformation of the fold and thrust belt continued to the north and south and was contemporaneous with Laramide deformation.

Available data on the timing of formation of Laramide structures indicate two phases of deformation, the early phase (about 74–65 Ma) characterized by more ductile behaviour of the crust and the late phase (about 60–45 Ma) characterized by more brittle behaviour. Structures of the early phase are widely distributed throughout the Colorado Plateau as well as along its margins, suggesting that stresses were widely distributed and evenly transmitted. Structures of the late phase, however, are restricted to a 200 km-wide belt parallel with the margins of the Plateau. During both phases of deformation, structures developed along the northern and eastern peripheries of the Plateau have orientations parallel to the Plateau margins. This suggests that the Plateau behaved as a rigid block and that strain occurred between it and the continental interior.

DISCUSSION

The coincidence in timing and spatial occurrence of several disparate tectonic events acts as a strong enticement to provide an explanation which links them by a common genesis. The explanation proposed here is an extension of Armstrong's (1974) insight and suggestion that a ductility contrast within the North American lithosphere was responsible for the localization and differing styles of deformation characteristic of the Sevier and Laramide orogenies. Armstrong attributed the ductility contrast to the precursor history of magmatism; crust in the area of high heat flow behind the active volcanoplutonic arc (Sevier belt) was ductile, whereas crust in the area formerly occupied by the magmatic gap (Laramide belt) was more brittle. The principal deficiency of Armstong's explanation is that it fails to account for the rapid eastward displacement of deformation (from Sevier to Laramide) and for the localization of Laramide structures within a narrow belt adjacent to the margins of the Colorado Plateau. Also, it does not address nor account for the episode of anomalous subsidence in the foreland, an observation which post-dated Armstrong's paper.

The timing and geographic occurrence of the major events previously discussed are shown in relation to the inferred history of plate interactions as a series of cartoons in Figs 8 and 9. These serve as an illustrative guide to the following proposal that genetically links these events by a single causal mechanism.

During the period of normal subduction, prior to about 80 Ma, décollement style deformation occurred along the Sevier fold and thrust belt in a region of high heat flow behind the Sierra Nevadan arc. Coeval subsidence of the foreland basin was confined to a relatively narrow belt adjacent to the foreland fold and thrust belt. Supracrustal loading by thrust sheets

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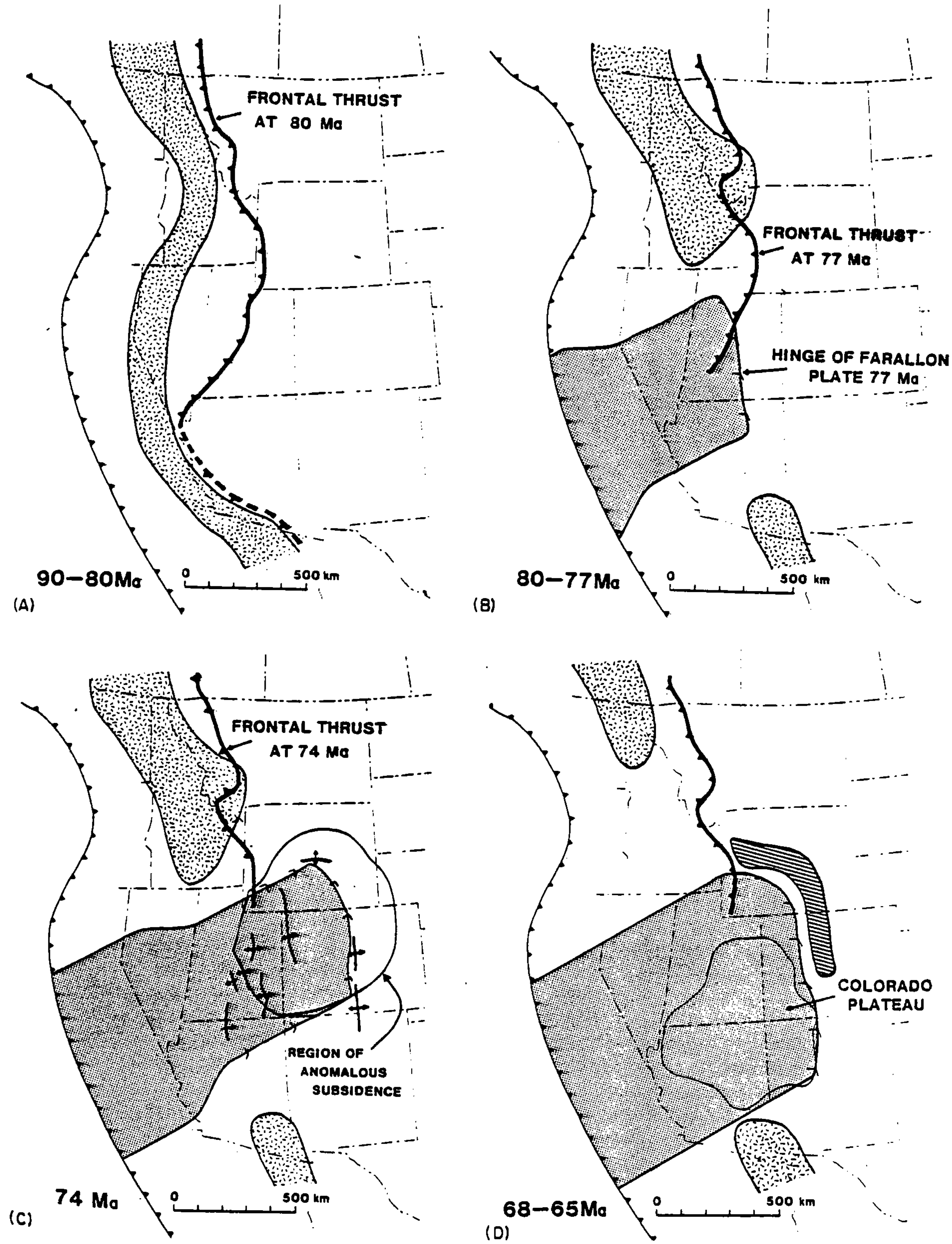

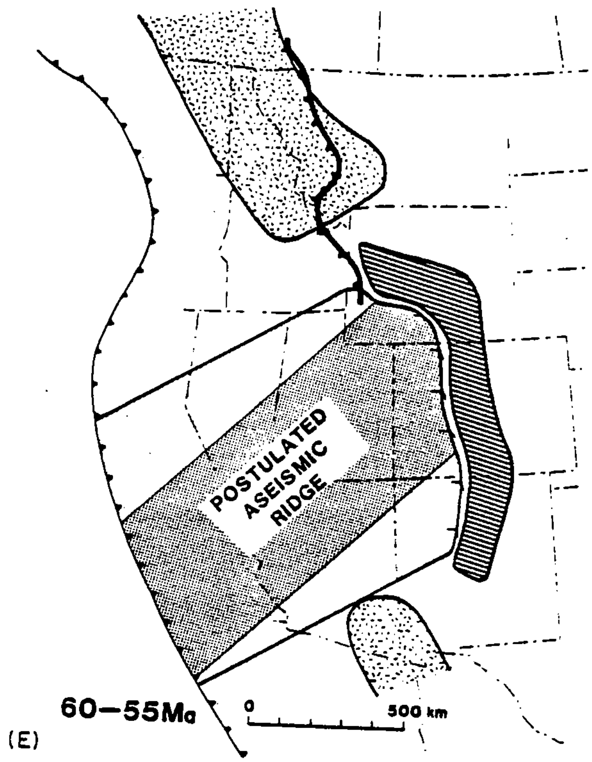

Fig. 8. Cartoons showing the inferred history of plate interactions, the evolution of subduction geometries, and the occurrence of the major tectonic and magmatic features discussed in the text. The inferred position of the subduction zone is indicated as in Fig. 1. The eastern deformational limit of the Sevier foreland fold and thrust belt is indicated by a thrust symbol with barbs on the west. The area occupied by Laramide deformation is indicated by the hachured fields and by the anticline symbols. Inferred position of subducted Farallon plate indicated by stipple pattern.

and derived sediment satisfactorily accounts for the location and amount of subsidence in the foreland basin.

Plate reconstructions indicate that North America changed direction and moved, in the hot spot reference frame, perpendicularly towards the trench by about 75 Ma. This change in the absolute motion of North America occurred during a period of rapid Farallon–North America relative convergence. An anticipated consequence of this change in absolute motion is that North America would override the trench and induce low-angle subduction. Inferences from geological data indicate that this episode of low-angle subduction was initiated before 75 Ma. Magmatism in the Sierra Nevadan arc ceased by 80 Ma, suggesting the termination of normal subduction and the initiation of low-angle subduction by then. In addition, if the genetic correlation of anomalous subsidence in the Colorado–Wyoming region with low-angle subduction is valid, the subducted Farallon plate must have reached Colorado by the mid-Campanian.

During the episode of low-angle subduction, western United States was directly underlain by the Farallon plate and a broad amagmatic area replaced the coastal volcanoplutonic arc. Heat flow in the region of the magmatic gap gradually dissipated, and the lithosphere cooled and became less ductile. Cessation of deformation along the Sevier foreland fold and thrust belt occured at about 75 Ma. At the same time, deformation within the same latitudinal zone was transferred to the broad region of the Colorado Plateau. The first phase of Laramide deformation may be attributed partially to a ductility contrast within the lithosphere of the North American plate, as suggested by Armstrong (1974). However, it is more likely that this effect was augmented by widespread transmission of stress into the overriding plate by the shallowly subducted Farallon plate. Broadly distributed open folds developed within the Colorado Plateau beginning at 74 Ma. The relatively ductile and widespread deformation during this period is inferred as the response to shear(?) coupling between the overriding and subducted plates and the concomitant transmission of stress over a broad area.

Toward the end of the first phase of Laramide deformation, from about 69 to 65 Ma, basement-cored, fault-bounded uplifts developed within a 200 km-wide belt along the north-eastern margin of the Colorado Plateau. Previously, this area had been amagmatic and, presumably, the crust was relatively cool and brittle. Localization of deformation within this narrow belt may be explained by substantial areal differences in the strength of continental lithosphere that were induced by the subducted plate. To the west of the early Laramide belt, under the Colorado Plateau region, North America was underpinned by the Farallon plate. To the east of this belt, the North American plate was underlain by asthenosphere. It is likely that the belt of early Laramide deformation was localized along the juncture between a double and a single thickness of lithosphere, that is, above the hingeline of the Farallon plate where it detached from North America and descended into the asthenosphere. This juncture would represent the position of greatest contrast in mechanical properties of the continental lithosphere and, correspondingly, the position of least lithospheric strength. Stress transmitted into the overriding plate by the Farallon plate would be relieved along the zone of least strength, coincident

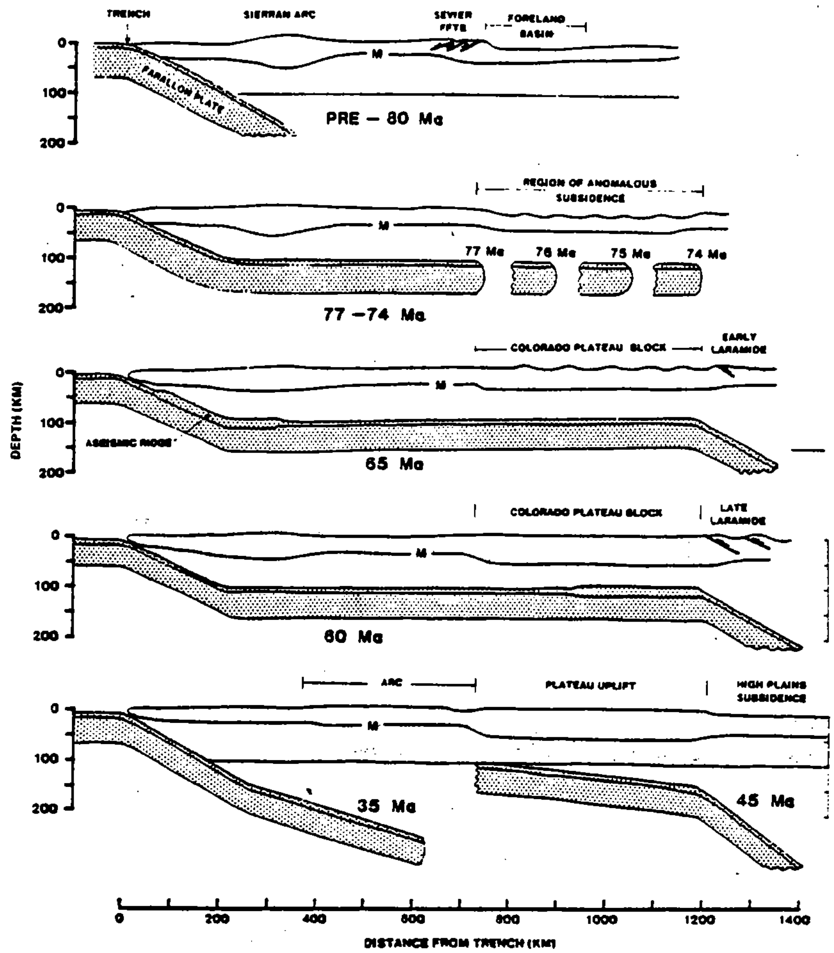

Fig. 9. Cross-sectional representation of Fig. 8 showing the inferred evolution of subduction geometries and the major tectonic and magmatic events considered to be genetic consequences of that evolution.

with the juncture between a single and a double thickness of lithosphere.

This suggestion also explains a long-standing enigma: 'Why did the Colorado Plateau behave as a tectonically rigid block, while regions to the north and south were intensely deformed?' The answer may be simply that the Plateau was underpinned and supported by the shallowly subducted Farallon plate. Because of the doubled lithospheric thickness, the Plateau behaved in a mechanically rigid fashion and intense deformation was restricted to that portion of the crust just beyond its margins.

As previously discussed, plate reconstructions suggest that an aseismic ridge had entered the subduction zone by, or more likely before, 65 Ma. The late Laramide episode of deformation may correspond to the time when the leading edge of the aseismic ridge reached the hingeline of the Farallon plate and began its descent into the asthenosphere. Presumably, lithosphere containing an aseismic ridge is more buoyant and, during low-angle subduction, would transmit more stress into the overlying lithosphere. In effect, the aseismic ridge would have 'bumped' the overlying lithosphere more intensely than before, but the strain still would have been concentrated along the zone of least strength. Increased transmission of stress by coupling between an aseismic ridge and the overriding plate also may account for the possible greater intensity of late Laramide deformation and its occurrence over a longer curvilinear belt. Although intellectually unsatisfying, no explanation for the pause between early and late Laramide deformation, if it existed, is given at this time.

The end of Laramide deformation at about 50 Ma corresponds to the time that the Farallon plate decoupled from North America. Because the Farallon plate no longer was in contact with overlying continental lithosphere, it no longer could generate and transmit stress into the North American plate. This decoupling event is inferred from the history of igneous activity in which the previously amagmatic area was filled gradually from the north and south by, respectively, southward and northward migration of magmatism. The source of these magmas is not established. It is possible that they were derived by partial melting of the detached Farallon plate as it descended into the asthenosphere. Alternatively, the Farallon plate may have broken, forming an eastern and a western segment, and renewed magmatism was a product of normal subduction of the western segment.

The preceding analysis and the proposed genetic relationship among several tectonic and magmatic events in the western United States is based upon two forms of argument. The first is their coincidental occurrence in time and space. The second involves a logic that links their development to the kinematics of plate interactions and the consequent evolution of subduction geometries. Although the expanations that have been advanced seem compelling at this time, they may be further strengthened through prediction of other events and subsequent verification of such predictions. To this end, two predictions are offered; verification has not been attempted and is beyond the scope of this report.

In analogy with the argument that low-angle subduction causes regional subsidence, decoupling should cause regional uplift of the area previously underpinned by the subducted plate. A potentially fruitful investigation would be to examine the evidence that, after 50 Ma, the Colorado Plateau and the area of Laramide deformation were sites of regional uplift. Some evidence that this occurred is summarized in reports by Mackin (1947), Epis *et al.* (1980) and Trimble (1980). A second prediction is that the lithosphere to the east of the regional uplift should have flexed downward in response to the supracrustal load induced by the uplift, and the High Plains to the east of the Laramide front should have become a site of deposition during the late Eocene and Oligocene. Again, some support for this prediction is found in summaries of Cenozoic events in the High Plains. A profound regional hiatus in the High Plains was developed during the Laramide orogeny. During this period, the High Plains were tectonically static and represented a surface of sediment transport by fluvial systems and widespread intense soil development. Thirty million years elapsed between deposition of the last sediments (late Maastrichtian) of the Cretaceous Interior Seaway and the next depositional event beginning in the late Eocene or early Oligocene. Renewed deposition during the middle Cenozoic may represent this predicted episode of lithospheric flexure, but a thorough analysis is warranted.

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