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LARAMIDE CRUSTAL THICKENING EVENT IN THE ROCKY MOUNTAIN FORELAND AND GREAT PLAINS

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Abstract. The Rocky Mountain foreland and Great Plains of the western United States were formerly part of a continental platform, adjusted by erosion and deposition in Cambrian through Jurassic time to near mean sea level, and therefore to a near-uniform crustal thickness of approximately 33 km. Today the region stands at regional elevations up to 2 km, isostatically supported by a crust exceeding 50 km in thickness. Reasonable estimates of Tertiary sedimentation and Laramide strain do not account for more than 15% of the implied thickening. However, from approximately 70-40 m.y. B.P., this region was underlain by a horizontally-subducting slab of Farallon plate lithosphere moving northeast; this slab could have been the cause of the thickening. Finite-difference thermal models with specified kinematics show only a temporary cooling of the base of the North American lithosphere by this slab. The excess weight of the slab would have depressed the region; plate-bending calculations show a quantitative agreement of predicted depression with upper Cretaceous isopachs. Since depression by this slab lasted until the Eocene at least, the latest-Cretaceous regression was probably caused by a Laramide crustal thickening event. The Farallon slab might have caused crustal thickening in two ways. Its excess weight would have drawn in ductile lower crust from surrounding regions. However, calculations show that this effect is too slow, too local, and too reversible to explain most of the crustal thickening. Therefore it seems likely that ductile lower crust was transported from SW to NE by shear stresses which the Farallon plate exerted on the base of the North American lithosphere. A preliminary finite-element calculation based on this hypothesis shows the correct general pattern of crustal thickening. An unexpected but encouraging result is that predicted principal compression directions are orthogonal to many Laramide basement uplifts.

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INTRODUCTION

Uncertainty and controversy over the basic mechanism of the Laramide orogeny go back well over 50 years. Today, the argument centers on the relative importance of vertical and horizontal movements. The "vertical-tectonics" or "upthrust" school holds that crustal shortening is secondary, that range-bounding faults steepen with depth, and that vertically directed forces displaced the basement blocks, draping and then tearing the sediments [e.g., Osterwald, 1961; Matthews, 1976; Stearns, 1978]. The "horizontal-tectonics" or "compression" school emphasizes shallow-angle thrust faulting (and subsidiary buckling) of the basement caused by horizontally directed compressive stress [e.g. Berg, 1962; Blackstone, 1963; Gries, 1983]. Recent geophysical evidence for thrusts dipping 30° under the Wind River range [Brewer et al., 1980] and the Laramie range [Brewer et al., 1982] are encouraging to the latter group, but it seems unlikely that thrust faults of comparable slip will be found under the Bighorns [Stearns and Stearns, 1978] or the Black Hills [Lisenbee, 1978]. More likely, it will turn out that the solution must encompass different styles in different ranges. A resolution of some of the controversy may be found through the study of wrench faulting, which is geometrically implied by the compression model [Sales, 1968; Stone, 1969]. The amount of strike-slip in Wyoming and Montana is still uncertain, but Chapin and Cather [1981] present strong evidence of major transcurrent motion in Colorado-New Mexico, which would bring the Colorado Plateau northward, or possibly northeastward [Hamilton, 1981].

Given this long controversy, it is suprising and probably significant that nearly all workers ascribe the fundamental cause of the orogeny to plate interactions in the West [Burchfiel, 1980]. Burchfiel and Davis [1975] attributed it to the warming and eastward slumping of the subduction-related volcanic/metamorphic belt. Livaccari et al. [1981] and Silver and Smith [1983] have suggested that an accretion event at the coast provided a stress pulse. But by far the most popular "underlying cause" is horizontal subduction of Farallon plate lithosphere from the coast to beneath the Rocky Mountain region.

Lipman et al. [1971] suggested a low-angle (second) slab beneath this region on the basis of K_2O/SiO_2 ratios in igneous rocks. Although this indicator has since been questioned [Meijer and Reagan, 1983], the basic idea of low-angle subduction found a receptive audience. Compilations of igneous dates began to show an eastward movement of the magmatic arc from 75 through 40-60 m.y. B.P. (depending on lattitude) and a less orderly return toward the coast in the Neogene [Snyder et al., 1976; Coney and Reynolds, 1977; Dickinson and Snyder, 1978; Keith, 1978]. Geophysicists contributed to the model by showing that low-angle subduction quickly becomes horizontal due to viscous stresses even if the slab is not buoyant [e.g. Tovish et al., 1978]; this model predicts a volcanic gap above the horizontal portion. Strong support came from present examples in Peru and central Chile/Argentina [Isacks and Barazangi, 1977]; both zones are amagmatic, and the Peru zone has low topography as predicted theoretically, while the Argentinian zone has foreland-style tectonics in the Sierra de Pampeanas [Jordan et al., 1983].

In reference to the Laramide, Lowell [1974] suggested that such a horizontal slab could be buoyant and assist in vertical uplifts. Cross and Pilger [1978] showed that it should be dense, and invoked it to explain the strange Campanian-Maastrichtian subsidence that took place well east of the Sevier orogen. Recently, Henderson et al. [1984] revived the idea of a bouyant slab by suggesting that it contained aseismic ridges comparable to the Hess and Shatsky Rises in the Pacific. Of couse, any unidirectional vertical force model of this sort will have trouble explaining both rising ranges and sinking basins in the same region. Sales [1968] and Dickinson and Snyder [1978] were among the first to advocate that shear stress between the slab and the North American lithosphere was responsible for horizontal compression, deformation, and differential uplift of the foreland.

Today, only 13 years after the inital suggestion, it is impossible to discuss the Laramide orogeny without assigning some role to this horizontal slab (Figure 1). A recent issue of Tectonics was devoted to the subject [Beck, 1984]. Henderson et al. [1984] suggested that subduction of one or two aseismic ridges on the Farallon plate may have been the cause of slab flattening. Jurdy [1984] and Engebretson et al. [1984] emphasized the spatial and temporal coincidence between the orogeny and rapid, near-orthogonal subduction. Finally, Wells et al. [1984] discussed the local effects of a probable Kula/Farallon/North American triple junction off the northwest coast.

Even if this model is accepted, the exact mechanism of deformation in the brittle upper crust remains uncertain. In this layer of rigid/plastic rheology, the locations of strain are greatly influenced by heterogeneities inherited from Precambrian orogenies [e.g. Cloos and Cloos, 1934; Chamberlin, 1945; LeMasurier, 1961; Hodgson, 1965; Bekkar, 1973; Allmendinger

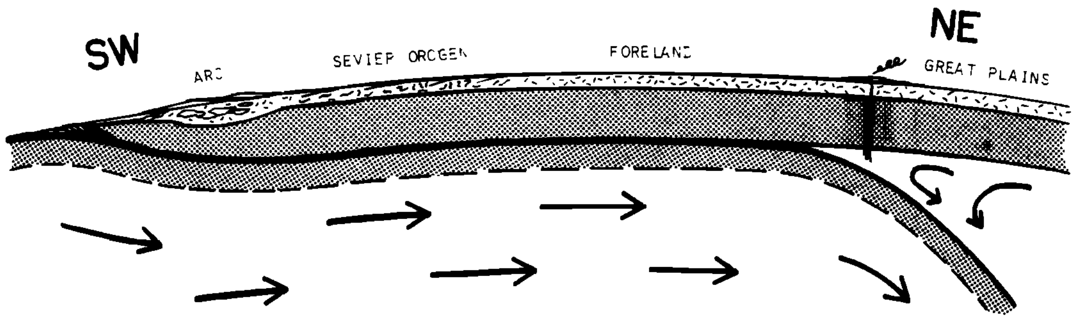

Fig. 1. Schematic cross section from SW to NE across the western continental margin, Sevier orogen, Rocky Mountain foreland, and Great Plains during the late Cretaceous-early Tertiary episode of horizontal subduction of the Farallon plate. Continental crust is stippled, oceanic crust is black, mantle lithosphere is shaded, and asthenosphere is left white. Inland volcanism was produced at a hingeline where the slab bent down and allowed water-rich metasediments on top of oceanic crust to contact hot asthenosphere.

et al., 1982]. Modeling attempts, which must assume a homogeneous continuum, are unlikely to reproduce all the diverse Laramide structures at once from a few equations and input parameters (although Sales [1968] was very successful in matching small regions individually). The contrast in size between the Farallon slab and the individual ranges is too great.

Perhaps computer modeling can advance the understanding of the Laramide orogeny in another way by investigating a huge but rarely-mentioned Laramide structure that formed in the ductile lower crust: the huge regional "root" that now supports the foreland and Great Plains at up to 2 km average elevation. The same shear tractions that Brewer et al. [1980] invoked to cause the Wind River range overthrust may also have dragged and transported ductile lower crust from within the Sevier orogen in the Southwest, and emplaced it under the foreland.

Assume, for the sake of argument, that weak layers almost completely decoupled the North American crust from its mantle lithosphere, and the North American lithosphere from the Farallon (Figure 2). If only one-tenth of the horizontal motion was transmitted across each decollement, very large strains would still result. Between 75 and 50 m.y. B.P., the Farallon plate moved some 3700 km with respect to North America [Engebretson, 1983]. Then (hypothetically) the mantle part of the North American lithosphere would have been dragged 370 km NE, and the upper crust about 37 km. The difference between the displacement of the top and the bottom of the North American crust (330 km) would produce a simple shear exceeding 10, and create a significant crustal thickening at the NE end of the region of contact. These figures are purely speculative, but serve to suggest that major changes in the structure of North America could have been caused by any degree of coupling to the mantle convection cell of which the Farallon plate was the surface expression.

In this paper, we will first examine what is known about the crustal thickness increase: the evidence for a Tertiary change; the argument in favor of an early Tertiary or Laramide date; five possible mechanisms of crustal thickening, and the reasons why three of them made only minor contributions. Then we will consider quantitative estimates of the location of horizontal subduction through time and its necessary consequences. These include heat tranfer between the plates, vertical loading of the North American lithosphere, plate flexure, surface subsidence, and induced lower crustal flow. Yet we will find that these consequences are insufficient to explain the Tertiary elevation history of the foreland. Lastly, we will consider the effects of horizontal coupling between the plates, and present a preliminary model that points toward a solution. If upheld by future calculations and tests, this model may be able to reconcile the old "vertical versus horizontal" tectonics controversy by providing an overall mechanism which could drive both local styles.

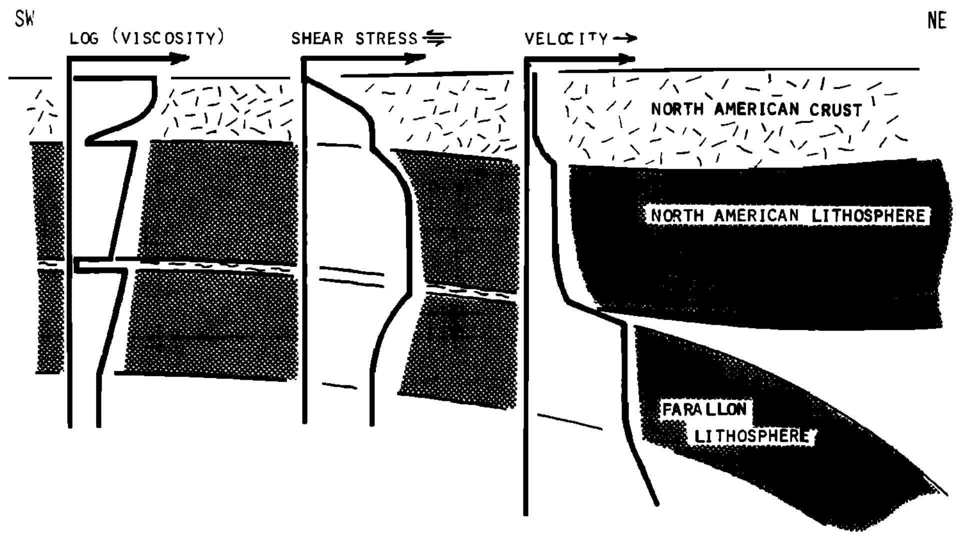

Fig. 2. Schematic diagram of the proposed mechanism of crustal thickening. Moving Farallon slab is largely (but not perfectly) decoupled from North American lithosphere. Horizontal shear stress, ultimately derived from mantle convection below, deforms North America in simple shear. Strain concentrates in the weak lower crust, and in the complex contact layer between the two plates. Both North American crust and mantle layers are thinned in the SW and thickened in the NE by this transport.

LARAMIDE CRUSTAL THICKENING EVENT

Evidence for Crustal Thickening

From the "mid-continent gravity high" on the east to the edge of the Great Basin on the west, the Great Plains and Rocky Mountain foreland are presently in overall isostatic equilibrium [Woollard, 1966]. On any east-west profile, average elevation increases almost linearly to the west, reaching 2 km in Colorado and Wyoming. Recent maps of crustal thickness (Figure 3) based on seismic refraction results show a corresponding increase in crustal thickness from 30-35 km at the Mississippi River to 50 km in Oklahoma, Kansas, Colorado, Nebraska, South Dakota, and parts of Montana [Soller et al., 1982; Allenby and Schnetzler, 1983]. The details of this increase are not well resolved by refraction data, as there are only two reversed refraction lines in all of the Great Plains states (North Dakota through Oklahoma). But Bouguer gravity data collected by Arvidson et al. [1982] become more negative in a nearly linear way from east to west, corresponding to the topography.

On the basis of Woollard's [1959] compilation of elevation versus crustal thickness worldwide, a 48 km crust should support 2 km of elevation. Thus there is no reason to invoke anomalously low densities in the mantle to explain the height of this region [e.g. Damon and Mauger, 1966; Damon, 1983]. In fact, Pn velocities are normal to fast (8.0-8.3 km/s; Allenby and Schnetzler [1983]) and heat flow varies from normal to mildly elevated (50-80 mW/m^2, Sass et al. [1976]) outside of the paleovolcanic area of the southern Rockies. So the present elevation is clearly not a result of transient high temperatures.

(In fact, there seems to be an effect of the opposite sense. In southeastern South Dakota and in Oklahoma where crustal thickness has been measured by refraction, it is "too great" to support the modest elevations there. In these cases, and perhaps generally in the Great Plains, the Bouguer gravity anomaly is less negative and the elevation lower than we normally associate with 50-km-thick crust. I believe this discrepancy can be explained as the result of Tertiary overthickening of the mantle part of the lithosphere, creating a more dense "antiroot" extending down into the asthenosphere. This would depress elevation and increase gravity. It would also explain the anomalously early arrivals of teleseismic P waves [Cleary and Hales, 1966] and S waves [Hales and Roberts, 1970] in the Great Plains and Mississippi valley. The model presented in this paper will suggest how such overthickening might have occurred.)

The stratigraphic record [Mallory, 1972] shows that the foreland region was a flat, quasi-stable platform close to sea level from the late Cambrian until the Jurassic (with the exception of an east-west belt shown in Figure 3 that was deformed in the Ancestral Rockies orogeny). During this long time there was no volcanism, little tectonism, and the positive thermal anomaly inherited from late Precambrian rifting gradually decayed [Armin and Mayer, 1983]. Thus it is very likely that the upper mantle was cool and stable in that time, and that the lesser elevation was due to a smaller crustal thickness (circa 33 km), made uniform and regulated by erosion and deposition. The record in eastern North America and elsewhere proves that sea level has not changed by anything like 2 km [Vail et al., 1977], so the regional crustal thickness must have increased since Jurassic time.

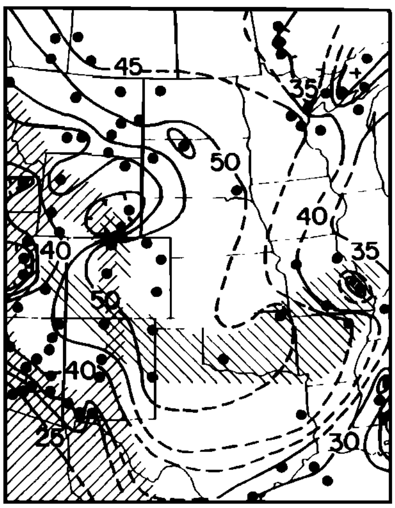

Fig. 3. Present crustal thickness in central North America, modified from Allenby and Schnetzler [1983, Figure 2]. Contour interval 5 km. Dots indicate locations of control points where thickness was measured by seismic refraction. Regions where crustal thickness may have been affected by late Paleozoic orogenies or Neogene extension are obscured by left- and right-handed crosshatching respectively. Crustal thickness in the remaining area is attributed to Laramide deformation.

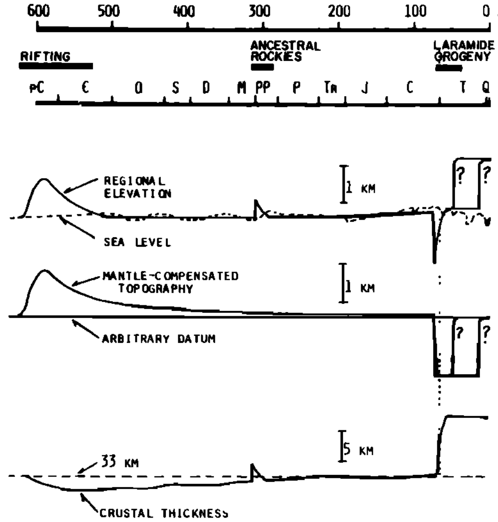

Fig. 4. Semi-quantitative histories of elevation and its compensation in the foreland. Only the last point of each curve and the crustal thickening by sedimentation are well controlled by data; the rest is a hypothesis. Since all compensation is assumed to be above 200 km, the top curve is the sum of the central curve and a multiple of the bottom curve. Paleozoic-Mesozoic sea levels are schematic, after Haun and Kent [1965]; Cretaceous-Tertiary levels are quantitative, after Bond [1978] and Vail et al. [1977].

Timing of the Event

The arguments which point to a Laramide age for this event are summarized in Figure 4. From paleo-elevation estimates plotted relative to changing sea level, we subtract topography compensated by the upper mantle events (slow cooling followed by the temporary intrusion of a horizontal Farallon slab) which will be computed below. The residual elevation is attributed to changing crustal thickness.

The slow subsidence due to cooling in the Paleozoic and Mesozoic [Armin and Mayer, 1983] did not result in major elevation changes; apparently the thickening of the lithosphere was balanced by small crustal thickness increases from net sedimentation. The Mesozoic Sevier orogen to the west loaded the foreland and flexed it downward anisostatically in the Cretaceous [Jordan, 1981] but this only affected areas near the edge of the fold/thrust belt. Rapid subsidence to the east in late Cretaceous time is most plausibly explained by the weight of the Farallon slab [Cross and Pilger, 1978].

Difficulties begin with the marine regression, which swept eastward in latest Cretaceous, lingered briefly in the Paleocene to form the shallow Cannonball Sea [Cvancara, 1976] and then removed the interior sea permanently. This regression occurred too early to be due to slab removal, which could not in any case explain a rebound to above the late Paleozoic elevation. Nor could it be due to the long-term trend toward lower sea level beginning at that time (-4 m/m.y. during 65-30 m.y. B.P.; Vail et al. [1977]), because the model calculations below predict that subsidence increased at more than 90 m/m.y. as the cold thermal anomaly under North America grew more intense. Finally, regression was probably not due to a massive influx of clastic sediments, since transitional units like the Fox Hills Sandstone are very thin in much of the foreland and Great Plains.

The two remaining possibilities to explain the regression are crustal thickening or the arrival of a buoyant region within the subducted Farallon plate [Henderson et al., 1984]. Of these, I prefer the crustal-thickening hypothesis because its effects are permanent, whereas the uplift caused by a feature like the Shatsky Rise passing beneath the foreland would only have lasted about 2 m.y. (based on an estimated width of 300 km and a subduction rate of 15 cm/yr). Considering the weakness of any alternative explanation for the latest-Cretaceous regression, it becomes the strongest evidence that the post-Jurassic crustal thickening was actually a Laramide event.

The succeeding Tertiary record is hard to interpret, as there are no absolute elevation indicators and sediment sources, climates, and base levels are all poorly constrained. Most likely, Eocene lake sedimentation was caused by disruption of drainage by mountain uplifts, rather than any regional subsidence [Chapin and Cather, 1981]. We know that most of the Wyoming ranges had been buried in molasse and/or volcanics by late Eocene, so that some were overtopped by Oligocene and Miocene sediments [e.g. Love, 1978]. This naturally suggests that the elevation may not have been great (Figure 4). Although excavation of the ranges was mostly post-Miocene, the absolute elevation history remains murky. Many early authors made a case for Neogene uplift on geomorphological grounds; but on the other hand Moore [1959] has shown that in many of these analyses pediments were misinterpreted as peneplains and assigned a regional significance which is inappropriate.

One extreme possibility is that the present regional elevation was established early, by the cumulative uplifts in the Paleocene, late Eocene, and Oligocene which have been inferred by Hallock [1933], Stagner [1939], Minick [1951], Moore [1959], Toots [1965], and others. In that case the Paleocene uplift could be due to crustal thickening, and the late Eocene uplift could be due to removal of the slab (Figure 4). Then, the Miocene-Quaternary development of relief could be ascribed to extensive normal-faulting [Love, 1970], increases in rainfall, accelerated erosion [Donnelly, 1982], and a 350 m drop in sea level [Vail et al., 1977].

Alternatively, it may be that there was a regional elevation increase in the Miocene or Pliocene. It would not follow that the crustal thickening event had two phases. Instead, the second uplift could be due to delamination of heavy mantle lithosphere from the base of the North American plate [Bird, 1979]. This sinking lithosphere could be either thickened North American lithosphere, or more likely some Farallon plate lithosphere left under the foreland and plains by Laramide subduction. It is characteristic of delamination to begin very slowly, go rapidly to completion, and produce about 1 km of uplift [Bird and Baumgardner, 1981]. Unfortunately, such events are very difficult to prove or disprove.

Because of this great uncertainty, it will probably be more productive to focus on explaining the amount and pattern of crustal thickening and net Tertiary uplift rather than its exact timing. The data available permit the hypothesis that crustal thickening occurred during the period of horizontal subduction.

Contributing Mechanisms

As shown in Figure 3, the excess crustal root in question extends from the overthrust belt on the west to the Mississippi River on the east, and from the

Canadian border on the north to Oklahoma on the south. For purposes of discussion, let us quantify the maximum crustal thickening as 51 - 33 = 18 km, and assume the crustal root has a wedge shape in E-W section, tapering to zero thickness 1000 km to the east. Then the average thickening in the region is 9 km. Several effects may have contributed to this:

1. Sedimentation. Crust added as sediment (including volcanics) should be counted only if it arrived during the Tertiary or Quaternary and remains today. Therefore, we need not consider how much Tertiary sediment may have already been removed from the intermountain basins and Great Plains. The amounts of sediment remaining can be estimated from the isopach atlases of Mallory [1972] and Cook and Bally [1975]. Using rough methods and leaning toward overestimation, I find that the volumes are probably no more than 2×10^5 km^3 of Paleocene, 2×10^5 km^3 of Eocene, 4×10^4 km^3 of Oligocene, 6×10^3 km^3 of Miocene, 6×10^3 km^3 of Pliocene, and 5×10^3 km^3 of Quaternary. If spread evenly over the whole region, this post-Cretaceous sediment would make a layer about 200 m thick, which is only 2% of the total crustal thickening. This figure would be lowered significantly (and would probably become negative) if we attempted to subtract the volumes of Paleozoic and Mesozoic rock eroded after the Cretaceous. Therefore, sedimentation was not a significant source of crustal thickening.

2. Intrusion. Addition of crust as volcanic deposits has already been considered under the general heading of sedimentation and found to be negligible. Likewise, the volume of Tertiary intrusives is negligible if it can be estimated by projecting their map patterns to the Moho to form vertical prisms. By this method intrusions would make up only 0.1% of the crust in the foreland. Therefore, vast plutons in the lower crust would have to be postulated to explain the crustal thickening through intrusion. Such plutons would have to extend far eastward from the furthest limit of known Tertiary volcanism across the Great Plains, and would imply a high present heat flow which is not observed. Of course, locally important thickening in the foreland cannot be ruled out.

3. Crustal shortening. Horizontal shortening of crust in the NE-SW direction would lead to vertical thickening if there were no compensating extension in the NW-SE direction. The horizontal shortening implied by "upthrust" models is on the order of 5% [Couples and Stearns, 1978]. Surprisingly, the shortening implied by the "overthrust" model is not much greater. Brewer et al. [1980] computed 21 km of overhang on their Wind River range section, and Gries [1983] estimated overhangs of other structures. But the net area of overhangs is only about 7% of the total area of Wyoming and northern Colorado which she studied. On the average, this amount of foreland shortening would produce 2300 m of regional crustal thickening, or only 13% of the amount required. A more serious objection is that thick crust is also found under the Great Plains, where there is no evidence for shortening.

4. Passive flow in the lower crust. Because crust must be "imported" by some means besides sediment transport or intrusion and the surface is insufficiently strained, it is natural to think in terms of independent mobility of a ductile, viscous lower crust. One theoretical possibility is that the horizontal pressure gradients around the edges of the region loaded by the Farallon slab could drive a converging flow (plane Poiseuille flow) of lower crust between the rigid layers of the brittle upper crust and mafic upper mantle. The rate of this process is extremely uncertain, because we are not sure of the rheology or even the composition of the lower crust. Therefore, I will show some calculations below that test whether the pattern of thickening would be correct, assuming a sufficient rate. In fact, it is not, because the transport of crust is entirely local (within state-sized regions).

5. Simple shear of lower crust. Although in the discussion of "passive flow" the Farallon slab was imagined to exert only a vertical load on North America, it should also have exerted a northeast-directed shear force. Dickinson and Snyder [1978] and Brewer et al. [1980] have appealed to this force to explain surface deformation; here I suggest that it may also have caused horizontal simple shear in the North American plate, with NE transport of lower crust (Figure 2). Calculations of stress between the plates in active oceanic and continental subduction zones give a shear stress of 20±10 MPa where the plates are in contact [Bird, 1978a, c]. This is about three times the shear stress which has effectively

detached the crust of the Zagros Mountains from underlying lithosphere [Bird, 1978b]. If stress of this order of magnitude was present over the vast horizontal contact surface between North America and the horizontal part of the Farallon slab, it could have had enormous consequences.

By the law of stress equilibrium, such a basal shear stress is converted into horizontal compressions and/or tensions in the upper plate which are greater by the ratio of slab width (or length, whichever is less) to lithosphere thickness. If the flat slab was 800 km long and wide, and the North American plate only 100 km thick, stresses on the order of 160 MPa would have been produced. This could help to explain structures like the Wind River thrust [Brewer et al., 1980].

The main virtue of this mechanism is that it predicts a net regional crustal thickening by "importing" crust from within the Sevier orogen in the southwest. This abstraction cannot be proven or disproven, since that area has been above sea level for some time (by unknown elevations) and has been extended (by an unknown fraction) in post-Oligocene time. However, the Sevier orogen is the most plausible source region in North America.

MODEL CALCULATIONS

Position of the Slab

Relative motion between the Farallon and North American plates has been calculated by Engebretson [Engebretson et al., 1981; Engebretson, 1983; Livaccari and Engebretson, 1983], by Carlson [1982], and by Jurdy [1984]. In Figure 5, I have used these vectors to project the edges of the flat slab inland past the truncated ends of the volcanic arcs as mapped by Lipman [1980]. Like the top boundary layer of any convection cell, the slab probably turned down eventually at a "hingeline." There it would have produced inland volcanism, as the hot asthenosphere came in contact with wet metasediments on the top surface of the slab, and either or both were melted. I have estimated the position of these hingelines as roughly parallel to the arc-front positions of Dickinson and Snyder [1978]. The area between the hingeline and the western continental margin was underlain by a horizontal slab of Farallon lithosphere traveling northeast.

Thermal Model

I have used the finite-difference program of Toksoz et al. [1971] as modified by Toksoz and Bird [1977] to calculate the thermal history in a representative cross-section cutting NE-SW across the Cordillera, foreland, and Great Plains and extending to 200 km depth. The inital condition in North America at 75 m.y. B.P. was a normal continental-platform geotherm [Bird, 1978c]. (This is clearly inaccurate in California and Nevada, but may serve as an adequate approximation in the quiescent foreland). The Farallon plate was assigned the theoretical geotherm of cooling ocean lithosphere where it enters at the southwest edge of the grid. The seafloor age used to compute this geotherm was taken from Engebretson [1983, p. 138]: it begins as 120 m.y. at 75 m.y. B.P. and then decreases. The critical thermal parameters assumed were a diffusivity of 0.012 cm^2/s, a subduction rate history taken from Engebretson [1983, p. 155], and a modest contact shear stress of 6 MPa. Results, exemplified by Figure 6, indicate that only part of the thickness of each plate is involved in the heat exchange because of the rapidity of subduction. North America is slightly cooled at its base, and the Farallon slab is warmed. The net effect is that a large negative thermal anomaly is introduced under North America. Its amplitude increases abruptly as the cold slab arrives, then increases slowly as North America is cooled, then decreases slowly as the Farallon plate becomes younger, and then decreases suddenly to a small residual when the flat slab eventually sinks away.

Plate Flexure

The Farallon slab imposed an increment of vertical load on North America. This is because the slab was much cooler than the asthenosphere it replaced, and therefore more dense. (Other density changes due to formation of oceanic crust, depletion of the upper mantle, and the basalt-eclogite phase change in that crust can be ignored to first order since ocean lithosphere is nearly a closed system.) The close contact between the Farallon and North American plates enforced by viscous lubrication effects [Tovish et al., 1978] would have been a much "stiffer" support than the convecting mantle below, so the

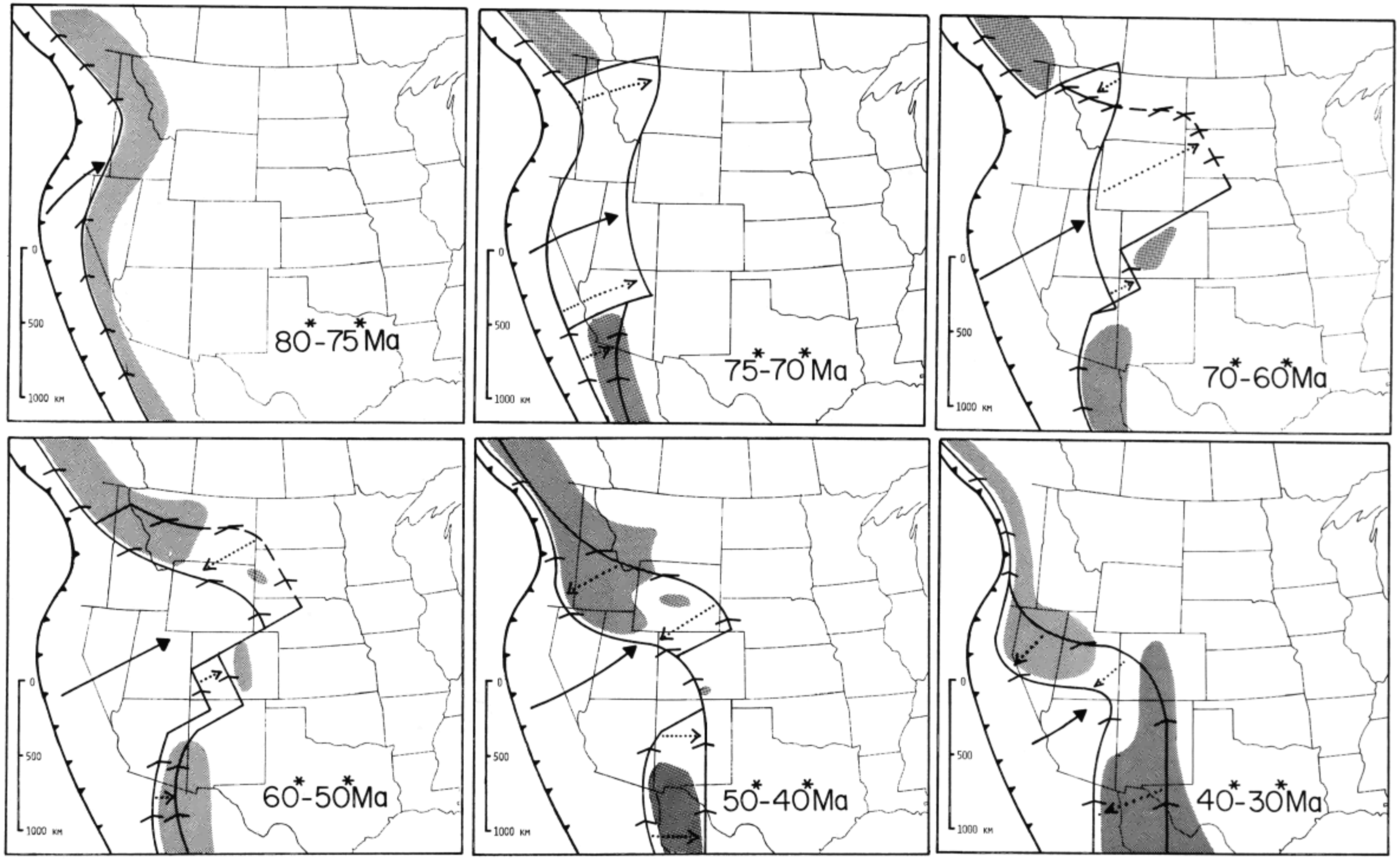

Fig. 5. Hypothesized locations of horizontal subduction beneath North America from late Cretaceous through Oligocene time. Palinspastic base map is from Hamilton [1978]. Solid arrows show velocity of the Farallon plate with respect to North America [Engebretson, 1983]; lengths are equal to 5 m.y. of relative displacement. Shading indicates regions of volcanism from Snyder et al. [1976] and Lipman [1980]. Solid lines without angle symbols at 70 m.y. represent edges of the flat slab, which should not have caused volcanism. Lines with dihedral angle symbols are suggested hingelines at the beginning and end of each period. Dotted arrows show the sense of hingeline migration. (*Note that all dates are based on cooling ages, so that if any significant time is required for intrusion and cooling, the slab may have actually reached the indicated position at an earlier time.)

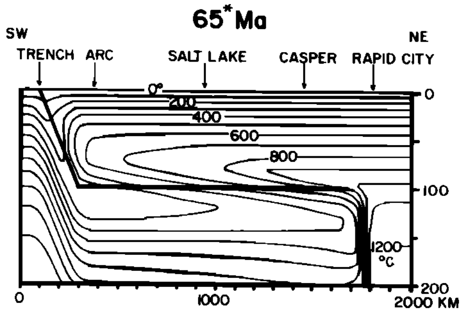

Fig. 6. One representative moment from a regional thermal history calculated by finite-difference methods in the cross-section of Figure 1. Vertical exaggeration of 5:1 for clarity. Slab motion was specified as descending to 100 km depth at 10° dip, then bending to horizontal. (The depth assumed for horizontal subduction is arbitrary and unimportant.) Initial geotherms assumed for the Farallon and North American plates can be seen at left and right margins, respectively. As horizontal subduction continues the Farallon plate will become younger and warmer, but the average temperature continues to drop through 55 m.y. B.P. as North America is cooled.

North American plate would have supported most of this excess mass. Of course, no absolute tension occurred at the plane of contact; the vertical normal stress was simply reduced below the value of vertical normal stress on the base of North America prevailing before horizontal subduction.

I applied the standard elastic plate bending equation to calculate the flexure of the North American lithosphere caused by this incremental vertical load. Bending is governed by

$$\frac{E\ T^3}{12\ (1-\nu^2)}\ \Delta^4 w = L + g(\rho_1-\rho_2)w$$

where E is Young's modulus, ν is Poisson's ratio, T is the thickness of the unrelaxed elastic part of the lithosphere, w is the vertical displacement, L is the vertical load stress, g is gravity, ρ_2 is the density of the asthenosphere below, and ρ_1 is the density of any loads filling the depression. This equation was solved by the Fourier-transform technique. For a reasonable effective elastic thickness (T = 40 km) and elastic modulus (E = 1 Mbar), the bending is confined to the edges of the slab (Figure 7). The interior region is nearly isostatic, with depression proportional to local load. Subsidence in the foreland at the critical time (65 m.y. B.P.) would have been about 3000 m assuming that the depression was filled with seawater.

This nicely confirms the qualitative argument of Cross and Pilger [1978] that a horizontal slab could explain the 2-3 km of deep-water Campanian-Maastrichtian sediments deposited in Wyoming and Colorado. Over most of the region the predicted subsidence is more than the amount deposited, allowing for paleodepths of up to 2000 m, which are consistent with the dominant black-shale lithology. An exact match of contour shapes is not to be expected because of such complications as uneven sediment supply, unfilled paleodepth, and additional subsidence due to the weight of sediment; yet it is encouraging that the predicted longitude of the zero contour (least affected) is about right. Note that a buoyant slab, as suggested by Henderson et al. [1984], would not have led to a successful prediction at this stage in the epeirogeny, although the subsequent arrival of a buoyant aseismic ridge embedded in the slab could have controlled the exact time of the subsequent regression.

Passive Crustal Flow

The vertically-integrated flux of plane Poiseuille flow in the lower crust should be proportional to the horizontal pressure gradient in the lower crust, which is roughly proportional to the surface slope. That is,

$$\int\vec{V}\ dz = \frac{-d^3}{12\ \eta}\ \vec{\nabla}P \simeq \frac{-g\ \rho\ d^3}{12\ \eta}\ \vec{\nabla}h$$

here $\vec{V}$ is horizontal velocity, z is the ertical coordinate, d is the thickness of he viscous layer, η is its dynamic viscosity, P is the pressure at a reference elevation, g is gravity, ρ is crustal density, and h is surface elevation. The resulting map pattern of crustal thickening (and thinning) rates can be determined from the convergence (divergence) of this vector flux; that is, thickening rate is proportional to the Laplacian derivitive of the topography. Because lower crustal viscosity is not well known, values cannot be attached to the contours which are presented in Figure 8. However, for reference it can be estimated that if the vis-

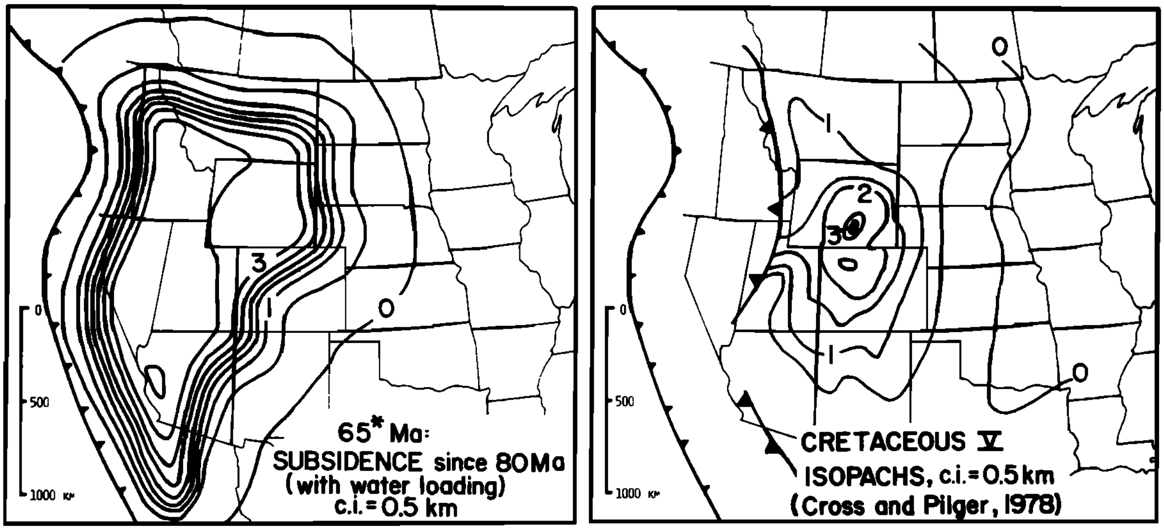

Fig. 7. North American plate flexure at the end of the Cretaceous. On left, theoretical prediction based on calculated vertical load of the Farallon slab and assumed infilling by seawater (since actual sediment fill was only a fraction of the total possible). On right, reconstructed Cretaceous V (Campanian-Maastrichtian) isopachs from Cross and Pilger [1978]. In the foreland and Great Plains, isopachs record a lower limit on the amount of subsidence because the region began near sea level. In the Sevier orogen west of the fold/thrust belt (thrust symbol) elevations may have been high enough initially so that subsidence did not cause transgression, and was not recorded.

cosity of the lowermost 10 km of the crust were uniform and Newtonian with the value of 10^{21} Poise, the contour interval shown would correspond to a geologically significant thickening rate increment of 10^{-2} cm/yr.

There are five reasons why I do not believe that this passive flow was a significant factor in the regional crustal thickening:

1. The pattern of predicted thickening is not right: it does not extend far enough to the east, its edges are too abrupt, and the central region is not thickened.

2. The predicted circumferential belt of thinned crust is not observed. Generally, the transport distances available with this mechanism are too small, so that a net regional thickening is hard to explain.

3. This model cannot explain regional elevation above sea level until after the slab has been removed. The timing is wrong; the interior sea largely vanished in latest Cretaceous, but horizontal subduction probably continued through mid-Eocene (Figure 5).

4. If the lower crust had a viscosity this low at shear stresses on the order of 1 MPa, then any reasonable nonlinear flow law would predict very low effective viscosities at higher stress levels. This great weakness would disagree, by about 2 orders of magnitude, with the empirical lower-crustal rheology determined from models of the Zagros orogeny [Bird, 1978b].

5. This crustal thickening mechanism is inherently reversible, and the excess crust collected over the slab would have begun to flow back after 40 m.y. B.P. when the slab was removed. Present elevation would then be only the result of hysteresis, so it would be necessary to postulate a much greater crustal thickness during Laramide times.

Basal Drag on North America

Dickinson and Synder [1978] presented a strong qualitative case that Laramide structures were created by horizontal drag forces from the Farallon plate, and concluded, "We close with the thought that the next step in the resolution of the

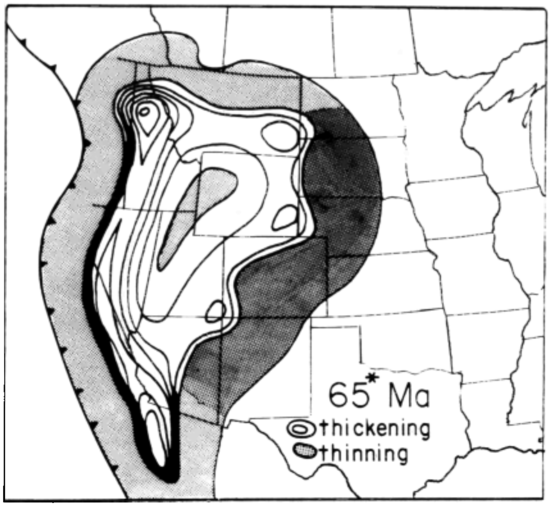

Fig. 8. Relative crustal thickening and thinning rates at the end of the Cretaceous, according to the mechanism of passive crustal flow. The quantity contoured is the Laplacian derivative of the vertical displacement from Figure 7. Contour interval unknown and dependent upon the effective viscosity profile in the lower crust. Note poor agreement with the pattern in Figure 3; deficiencies of this model are discussed in the text.

classic Laramide problem must be an analysis of the mechanical behavior of a plate of lithosphere given the ... boundary conditions inferred here." This analysis will be very complex, because the problem is inherently three-dimensional. In vertical section, it involves three strong layers (North American upper crust, North American mantle lithosphere, and the Farallon slab) divided by two weak layers (North American lower crust and the subducted sediments). The rheology is uncertain in all of these layers. Furthermore, other complexities develop with finite strain: heat is advected with the rocks and changes the geotherm, while crustal thickening leads to uplift which introduces gravitational stresses. There is a danger that the essential concept will be obscured by these complications. Therefore in this paper I will make a large number of simplifying assumptions and approximations in order to present a single illustrative model of the greatest possible generality.

First, the deformation of the Farallon slab can never directly be constrained by data and must remain speculative, so there is little value in modeling it. I have treated the slab merely as a source of shear stress on the base of North America, with the direction of stress determined by the kinematic plate model. The stress is considered to be uniform over the area southwest of the hingeline, where there is close contact, and zero outside.

Second, the upper crust can be approximated as rigid, since foreland strains are only 5-10% and the Great Plains are undeformed. In comparison, the strains required in the lower crust and lithosphere are 10 to 100 times greater, and will only be slightly affected by yielding at the surface. With this approximation, the model now involves only deformation of the North American lithosphere and the boundary layer of lower crust which it drags along with it. This deformation is controlled by driving shears from below, rigid boundary conditions, and restraining drag from above where the ductile lower crust is sheared. A continuum of models is possible, depending on whether the resistance to the deformation of the lithosphere comes mainly from its own internal strength or from crustal drag. The results are intuitively obvious for the case where the lithosphere has no strength (its motion imitates that of the slab). In the other end-member, the lower crust is weak (a necessary condition for the proposed mechanism) and its resistance is negligible compared to the resistance provided by strength in the lithosphere. Crustal thickening rate is proportional to the thickening rate of the lithosphere layer, because the average horizontal velocity of the lower crust is a uniform fraction of the horizontal velocity of the lithosphere. This is the case presented here.

Third, I have modeled only the beginning of deformation, so that it is possible to ignore topographic stresses, variations of the geotherm, and variations in lithosphere thickness. Consistent with this, the lithosphere is assumed homogeneous and isotropic, and assumed to deform according to the nonlinear cubic-creep dislocation-climb flow law [Weertman and Weertman, 1975]. This has previously been found to give an adequate representation of continental deformation in the case of the Himalayan continental collision [England and McKenzie, 1982].

Note that the problem is now nondimensional, in the sense that it has not been necessary to assume any numerical values of rock strength or basal drag. Therefore, general patterns of stress and velocity and thickening rate can be calculated, although their absolute magnitudes remain uncertain.

I have limited the model domain to the foreland and Great Plains, and assumed that the hot and mobile crust to the southwest is so weak [Burchfiel and Davis, 1975] that it provides one deviatoric-stress-free boundary on the west side of the foreland. In future models, it may be desirable to use a constant-pressure rather than a stress-free condition on this side, to represent the spreading tendency of the elevated Sevier orogen. This is not done here, because it would destroy the generality of the model by introducing a second deforming force. In justification, it can be noted that this second force was active long before the interval of flat subduction, without any apparent effect on the foreland. The other three boundaries, facing adjacent North American craton, were held fixed (Figure 9). The area of North America underlain by Farallon slab at shallow depth was acted on by a uniform shear traction; outside of this area, the base was free of shear stress.

The calculation was performed with the finite element method of Bird and Piper [1980], modified only slightly to allow for basal drag [Bird and Baumgardner, 1984]. The essential feature of this method is that the stress equilibrium equations (and the strength of the lithosphere) are integrated vertically at each point to create a "membrane-tectonic" model. That is, we solve the differential equations

$$\frac{\partial}{\partial x} \int \sigma_{xx}\, dz + \frac{\partial}{\partial y} \int \sigma_{xy}\, dz + \sigma_{xz}(z_b) = 0$$

$$\frac{\partial}{\partial y} \int \sigma_{yy}\, dz + \frac{\partial}{\partial x} \int \sigma_{xy}\, dz + \sigma_{yz}(z_b) = 0$$

involving stress σ_{ij} for the stress and velocity subject to an assumed stress/strain-rate relation:

$$\dot{e}_{ij} = \frac{(\sigma_{ij} - \delta_{ij}\, P)}{2\eta}; \; \eta = f(\dot{e}_{ij}, T, P)$$

where $\dot{e}_{ij}$ is strain-rate, δ_{ij} is the Kronecker delta, and T is temperature. The nondimensional output of this very simple model is presented in Figure 9. To apply these results to the Laramide orogeny it is only necessary to scale and rotate the model (Figure 10). The position and direction of motion (N60°E) illustrated in Figure 10 are appropriate for a time about 54 m.y. B.P., near the Paleocene/Eocene boundary. Each feature of the output corresponds well to some aspect of the Laramide orogeny:

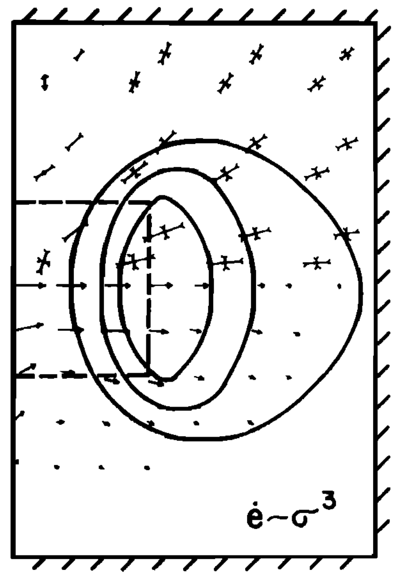

Fig. 9. Results of a finite element calculation of deformation in a plate of non-Newtonian lithosphere. Left boundary is free; others are fixed. Uniform shear traction to the right is applied to the base within the small dashed rectangle; other top and bottom boundaries are free. Since solution has a mirror plane of symmetry, stress is shown only in the top half, and velocity only in the bottom half. Contours show the rate of thickening of the plate, proportional to the convergence of the velocities. All magnitudes are relative, and depend on the strength of the plate and/or the size of the basal shear stress.

1. Maximum velocity occurs over the center of the slab and in the direction of the basal shear (N60°E). As velocity decreases in three directions (NW, NE, SE) outward from the center, local rotation is implied. The southeastern half of the computed velocity field in Figure 9 corresponds nicely to Hamilton's [1981] model of clockwise rotation of the Colorado Plateau about an Euler pole in Texas. This model suggests that the rotation in the crust may have been driven by a similar (perhaps faster) rotation of the mantle

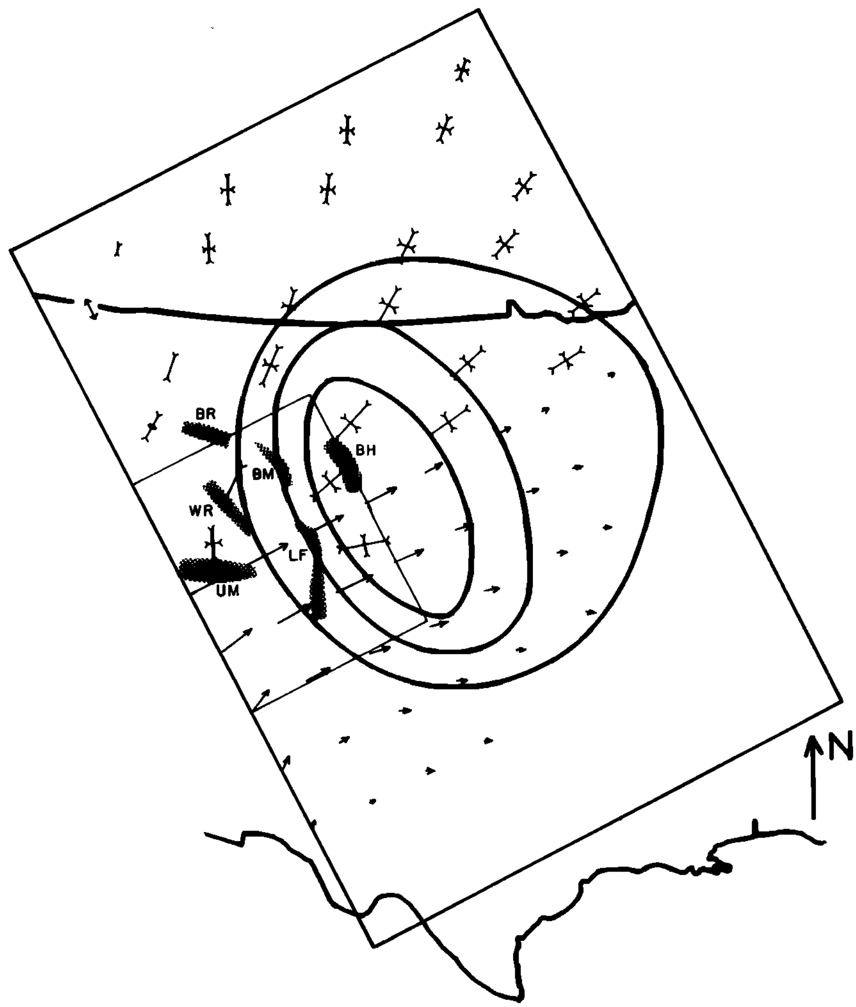

Fig. 10. Application of the model from figure 9 to the Laramide deformation of North America. The model is rotated and scaled so that the small rectangle approximately matches the location of the Farallon slab under the foreland at the Paleocene/Eocene boundary and basal shear stress is exerted in the direction of slab motion (N60°E). The Sevier orogen is assumed mobile and weak and in this analogy only serves as a "free" boundary on the foreland. Contours now represent the rate of crustal thickening (compare to Figure 3). A number of major Laramide uplifts are shown with shading and identified below; note that predicted directions of principal compression are nearly perpendicular in each case. Precambrian basement uplifts illustrated in this figure are BR, Beartooth Range; BM, Bighorn Mts.; BH, Black Hills; WR, Wind River Range; UM, Uinta Mts.; LF, Laramie and Front Ranges.

part of the North American lithosphere.

2. The predicted pattern of crustal thickening rates is in reasonable agreement with the present map of crustal thickness (compare Figures 10 and 3). The region of thickening extends eastward to the Mississippi River and northward to the Canadian border, as does the present thick crust. To the south, it terminates around the northernmost part of Texas, which may be too far north; but a fully time-dependent solution would include a migration of the flat slab from north to south (Figure 5) which might well correct this discrepancy.

3. The predicted directions of great-

est horizontal compression in the model plate are locally orthogonal to many of the major basement uplifts in the foreland (Figure 10). Some amount of horizontal compression normal to strike is required according to either the "horizontal compression" or the "vertical tectonics" models of basement uplift [Couples and Stearns, 1978], so this is a very encouraging result. Perhaps in some cases (Wind River range) this stress was the fundamental cause of uplift, while in others (Black Hills) it was an uneven crustal thickening which provided a vertical impetus, and a weaker horizontal compression that controlled the trend of the structures produced.

It remains to be seen whether the fortunate results of this simplified nondimensional model will be preserved in a fully three-dimensional treatment. Advection of heat, the force of gravity on elevated regions, the changing strength of the thickening and thinning layers, and the changing position and direction of the Farallon slab must all be considered in making a realistic and complete model of the orogeny. But it seems inescapable that in some way the horizontal tractions exerted by the Farallon plate during this episode of horizontal subduction have had a major effect in the shaping of the crust of central North America.

CONCLUSIONS

1. Crustal thickening by up to 20 km has occurred in the Rocky Mountain foreland since Mesozoic time, with lesser amounts of simultaneous thickening in the Great Plains. The regional elevation of these provinces is now isostatically supported by this thick crust.
2. Simple calculations show that sedimentation, intrusion, and crustal shortening can account for no more than 15% of the thickening.
3. Crustal thickening appears to coincide in space and time with horizontal subduction of the Farallon plate beneath North America, as outlined by patterns of inland volcanism.
4. This thickening began, and may have been completed, during the Laramide orogeny. Crustal thickening was a more significant controlling factor on the late Cretaceous-Paleocene regression than either the subduction of buoyant plateaus or the relatively small global change in sea level.
5. Theoretical models of such horizontal subduction show that the Farallon slab is more important as a source of forces on North America than for its thermal effects.
6. Calculations of lithosphere bending caused by the excess weight of this slab quantitatively confirm the concept of Cross and Pilger [1978], that horizontal subduction caused the late Cretaceous subsidence in the central foreland.
7. The vertical force on North America would have caused some amount of crustal thickening through passive flow of the ductile lower crust. However, it is doubtful that the lower crust was weak enough to make this effect significant. Further, calculations of the pattern of thickening show very poor agreement with present elevation and crustal thickness patterns.
8. A very simplified and preliminary calculation suggests that horizontal drag forces on the base of the North American plate exerted by the Farallon plate can explain both the pattern of crustal thickening and the pattern of deviatoric stress prevailing in the Laramide orogeny.

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