To: Greg Davis

FORMATION OF THE ROCKY MOUNTAINS, WESTERN UNITED STATES: A CONTINUUM COMPUTER MODEL

Peter Bird

Professor of Geophysics & Geology

Department of Earth & Space Sciences

University of California

Los Angeles, CA 90024

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At last, I've finally "put laser to paper"!

Best wishes,

P. Bird

ABSTRACT

One hypothesis for the formation of the Rocky Mountain structures in late Cretaceous through Eocene time is that a plate of oceanic lithosphere subducted along the base of the North American lithosphere. The movements of this plate have previously been determined from paleomagnetism, and the flat part of the slab has been traced using reconstructed patterns of volcanism. Now, new techniques of finite-element modeling allow prediction of the thermal and mechanical effects on the North American plate. Using a realistic temperature-dependent rheology and a simple plane-layered initial condition, the consequences of horizontal subduction in the time interval 75 to 30 m.y.B.P. have been computed. In this model, the mantle part of the North American plate was detached from the upper crust, across a ductile shear zone in the lower crust, and expelled northeastward from the region of flat subduction. The moving oceanic plate then deformed the crust by both pure shear (at the top) and simple shear (at the base). When horizontal subduction ceased, this crust was exposed to hot asthenosphere, uplifted, and weakened; extensional strain then began spontaneously. Successful predictions of this model include (i) the location, timing, amount, and direction of horizontal shortening inferred from Laramide structures in Montana through New Mexico; (ii) massive transport of lower crust from southwest to northeast, resulting in thickening by up to 30 km in the Rockies; (iii) the location and timing of the subsequent extension in early metamorphic core complexes and the Rio Grande rift; and (iv) the total area eventually involved in Basin-and-Range style extension.

INTRODUCTION

The formation of the Rocky Mountains, in an event known as the Laramide Orogeny, has been studied by geologists for a century. It is well known that deformation began in the late Cretaceous epoch, peaked in the Paleocene, and waned in the Eocene. All mountain ranges (and many smaller structures studied during petroleum exploration) show evidence of minor horizontal shortening of the crust in the E-W or NE-SW direction. Although there has been much controversy (1) over the exact mechanism of basement uplifts, workers of different schools would probably agree that net shortening in Wyoming was roughly 5% (2) to 10% (3). This is not enough to explain the thickening of the crust from about 33 km to over 55 km which also occurred during the Laramide Orogeny (4), and which currently supports the high regional elevation. Therefore, some mechanism involving larger strains in the ductile lower crust must be inferred. One additional complication is that most of the Laramide structures were depressed and sediment-covered during their formation, only rising to become topographic features after the end of Laramide time (5).

The concept that the Rockies might have been formed by "subcrustal convection currents" goes back at least to Longwell (6), who applied concepts of Holmes and Griggs to this region. Now, the modern theory of plate tectonics has made it possible to demonstrate that vast areas of oceanic crust and lithosphere were subducted (underthrust) beneath the west coast of North America during this time (7, 8). A link between these two processes was suggested by Dickinson and Snyder (9), who proposed that the subduction was horizontal, with the oceanic lithosphere sliding along the base of North America as far inland as the Black Hills. They suggested that the contact shear stresses caused the shortening strains seen at the surface. Horizontal subduction has also been invoked to explain the sudden subsidence of the region in the late Cretaceous (10). Finally, I have previously suggested (4) that horizontal subduction transferred crustal material into the Rockies region from the southwest, increasing the crustal thickness by about 65%.

The purpose of this paper is to test whether the hypothesis of flat subduction is tenable by a quantitative prediction of its effects. This has been done with a set of new finite-element techniques developed especially for this problem (11) which permit the calculation of crust and mantle-lithosphere displacement, thickness, and temperature through time. The boundary conditions of the simulation are obtained from plate tectonic theory and are reasonably certain.

This model does not have the spatial resolution to predict individual structures like the Wind River or Big Horn ranges, but it does consistently predict a curved belt of crustal compression with the correct location, timing, and orientation. (Only the amplitude seems to be adjustable, through the assumed rheology.) Furthermore, the model predicts the transfer of lower crustal material into the Rockies region from the coastal region in a great "solitary wave". After the eventual removal of anomalous masses in the mantle, this extra crust would cause a buoyant uplift, explaining the present elevations. Another, unexpected feature of these solutions is that the mantle layer of the North American lithosphere is entirely stripped away from the region west of the Rockies. This provides a simple unifying explanation for the cause, location, and timing of the extensional strain that followed the Laramide Orogeny, and created the Basin-and-Range province and the Rio Grande Rift.

Thus, the hypothesis of horizontal subduction is doubly validated: it explains the mechanism of the Laramide Orogeny, and it also suggests an explanation for the next orogeny, which has commonly been considered a distinct and unrelated event.

Computational Methods

The complete derivation of model equations is given in reference (11). The problem is difficult because the strength of rocks is exponentially dependent on temperature (above some transition temperature), while temperature gradients are large and depend on the deformation. The duration of this model (45 m.y.) is such that neither an adiabatic nor a steady-state approximation of temperatures is adequate. Thus, the calculation proceeds through timesteps (of 1 m.y.), each of which is artificially partitioned into an adiabatic translation and a static period of heat diffusion.

The horizontal components of velocity are computed by the "thin-plate" finite element method (12, 13, 14). The model domain is the lithosphere of North America, which is very much wider (5000 km) than it is thick (100 km). Therefore, it is adequate to solve the stress-equilibrium equations in vertically-integrated form, effectively condensing the strength of the plate into a single plane. Vertical force equilibrium is represented by the isostatic approximation, which is that each column floats on a fluid support at an elevation determined by its density structure. This method reduces the finite-element grid from 3 dimensions to 2 (Figure 1), with an indispensable reduction in cost. The finite element grid deforms over time to follow the path of the more rigid material at the top of each layer. One innovation in this project is the use of two stacked grids (initially identical) to represent the crust and mantle layers of the North American lithosphere respectively; this allows their velocities to differ, which is an essential feature of the model. The vertical integration of strengths is performed numerically at each of 7 Gauss integration points in each finite element. A simultaneous vertical integration of compliance (inverse of effective viscosity) provides the coefficients necessary to compute the shear stresses exchanged by the layers when their horizontal velocities differ.

The rate of thickening of each layer is determined using the assumption of incompressibility. One effect causing thickness changes is divergence, in the horizontal plane, of the horizontal velocity of the grid and the strong upper part of each layer. This will be referred to as "pure shear" since it is irrotational in any vertical cross-section.

A second effect is divergence, in the horizontal plane, of the relative flux of ductile material that is involved in "simple shear" to accommodate different horizontal velocities of layers. This relative flux (relative to the moving top of the layer) depends on the horizontal velocities of the grids, and also on the vertical temperature gradient, which determines the thickness of the simple-shear boundary layer. (Wherever crust is transported by simple shear, it remains a part of the crustal layer, and is never dragged down below mantle material; in nature this would be enforced by its relative buoyancy.)

The third effect is independent flow of ductile material in response to lateral pressure gradients associated with topography. In the crust, this effect tends to smooth the thickness through a nonlinear "diffusion" of crustal thickness. Like displacement, layer thickness is assumed to be laterally continuous, and is parameterized by quadratic finite-element basis functions.

In the computation of temperatures, lateral heat conduction is neglected. At each of the Gauss integration points of each of the elements, the geotherm (temperature as a function of depth) is represented as the sum of steady-state quadratic functions and a set of 5 decaying eigenfunctions of the homogeneous one-dimensional diffusion equation. Wherever simple shear between layers would tend to cause a temperature discontinuity at the base of a layer, this is corrected by adding an appropriate z^3 term to its geotherm (where z is depth).

Convergence tests documented in (11) suggest that the results presented in this paper have a relative precision (RMS error/RMS signal) of about 15% for the variables of crust and mantle displacement and mantle thickness. However, the changes in crustal thickness have only about 50% RMS relative precision, and should be considered as suggestive rather than definitive. Greater precision is not presently attainable, as each run already consumes about 2 hours on either a Cray X-MP/48 (1 processor) or an IBM 3090/VF. The main reason for this expense is that the nonlinearity of the rheology requires the velocity solution to be iterated about 14 times per timestep.

Rheology of North America

In this problem the total strains of interest range from 0.05 up to 1000, so the contribution of elastic strain is negligible. Deformation is governed by one of three laws based on laboratory experiments, each of which places an upper limit on the shear stress σ_s. At low temperature, the important limit is set by frictional sliding on any and all planes:

$$\sigma_{\mathrm{s}} \leq \mu(\sigma_{\mathrm{n}} - P_{\mathrm{H_2O}}) \tag{1}$$

where μ is the coefficient of friction, σ_{n} is the (positive) normal stress on the same plane, and $P_{\mathrm{H_2O}}$ is the pressure of water in pores (assumed hydrostatic). At higher temperatures, the relevant limit is set by thermally-activated, power-law dislocation creep:

$$\sigma_{\mathrm{s}} \leq 2\alpha\left(2\sqrt{-\dot{e}_1\dot{e}_2 - \dot{e}_2\dot{e}_3 - \dot{e}_3\dot{e}_1}\right)^{(1-n)/n} \dot{e}_{\mathrm{s}} \exp\left(\frac{\beta + \varepsilon P}{T}\right) \tag{2}$$

where $\dot{e}_i$ are the principal strain rates, $\dot{e}_s$ is the shear strain rate (on the same plane as σ_s), P is total pressure, T is absolute temperature, and α, β, ε, and n are empirical constants. (Values of these and other parameters are contained in Table 1.) The third limit is a plasticity condition which is independent of P and T:

$$\sigma_{\mathrm{s}} \leq \sigma_{\mathrm{p}} \tag{3}$$

In applying all of these laws, the vertical stress σ_{zz} is assumed to be lithostatic. To avoid numerical difficulties, an upper limit η_{max} is also placed on the effective viscosity at all points.

It should be noted that two of these parameters (α and β of the crust) where adjusted to values that gave the proper magnitude of Laramide surface strain. However, these values have to be taken in the context of many other assumed parameter values (especially thermal parameters) and are not presented as experimental determinations. The question of how this modeling project can constrain the rheology of North America will be addressed in a later paper.

Initial Conditions

The model begins at 75 m.y.B.P. in the late Cretaceous epoch. At this time, the future Rocky Mountain region and Great Plains were close to sealevel, and had been since the late Cambrian epoch (16). Therefore, I assume that heat diffusion had established a uniform steady- state geotherm with a typical platform heatflow of 54 mW m^{-2}, and that the combination of isostasy, erosion, and deposition had leveled the crust to a uniform thickness of 33 km. The initial value of 70 km chosen for the mantle lithosphere is somewhat arbitrary, since the lithosphere/asthenosphere boundary is gradational. Because the strength of this layer is concentrated at the top, the lower limit of integration makes little difference.

Further west, within about 600 km of the coast, there was a mountain belt comparable to the present Andes (17). This had developed during the previous 80 m.y. above a subduction zone of normal geometry, and its eastward spreading under the force of gravity was responsible for the Cretaceous Sevier Orogeny that formed the Overthrust Belt (*ibid*). Initially, I attempted to include such a cordillera in the initial condition. But it was soon apparent that the balance of forces in such a system is subtle and not easily guessed, for I could not find an initial condition that would be in steady-state. This level of realism is deferred to future studies, and the reader is simply cautioned that crustal thicknesses were originally greater to the west of the Rockies than in the model presented.

Boundary Conditions

The northern, eastern, and southern margins of the finite-element grid (Fig. 1) are fixed. Boundaries in Mexico and Canada are placed where Laramide strain becomes secondary to the effects of other Tertiary orogenies which are not modeled. The eastern and Gulf-coast boundaries are natural; the strength of oceanic lithosphere is so much greater than that of continental that strain will be negligible in the former. The western edge of the continent was the upper plate of a subduction zone, so this edge was tapered and does not require a boundary condition.

The bottom boundary is divided into two areas: the area overlying normal hot asthenosphere, and the area overlying flat-subducting oceanic plate. The boundary between these areas moved inland and then westward again during the interval modeled (Figure 2), and its progress may be tracked by the changing patterns of volcanism in the western United States (4, 9).

Where the model overlies asthenosphere, its basal temperature is fixed (by convection) at 1440 K. Mechanically, it is supported on a perfect fluid (with density 3210 kg m^{-3}) in which there are no shear stresses. However, where this asthenosphere comes into contact with the crust, I assume that crustal melting and/or convection will limit the temperature to 1073 K.

To determine the excess mass of the oceanic slab, I first reconstruct the Farallon plate of oceanic lithosphere, using finite rotations from Engebretson *et al* (8). I then apply the boundary-layer cooling model by assuming that its vertically-integrated excess weight with respect to asthenosphere is 6600 Pa $\times \sqrt{t}$, where t was its paleoage or cooling time, in years. The normal fluid support pressure on the bottom boundary is reduced by this amount, so that after isostatic adjustment the surface of the continent is depressed by about 2-3 km above the slab (4, 10). The horizontal velocities of the slab are obtained from rotation poles of Engebretson *et al* (8), and used as velocity boundary conditions over most of the area contacted. Note that velocity is imposed on the weak base of the bottom layer, and the velocity of its strong upper levels and its finite-element grid may be less, depending on rheology and temperature. Also, the shear stress necessary to impose this velocity boundary condition is monitored, and if it exceeds 10^8 Pa then that shear stress is substituted (to simulate formation of a ductile fault). The appropriate value of this stress limit is the most uncertain of the model parameters.

Where this slab makes contact with the base of North America, a temperature boundary condition of 700 K is imposed. This value is the average obtained in a two-dimensional finite-difference model of heat advection and diffusion (including the oceanic plate in the domain) that was conducted separately for a representative cross-section. Of course, a more realistic boundary condition will be needed before final estimates of the optimal plate rheology can be determined. Note that this temperature is lower than the initial temperature at the base of either crust or mantle, so that the thermal effect of the slab is to cool and strengthen the parts of the continent that it touches.

Summary of the Model

As the flat-subducting slab of oceanic lithosphere began to slide along the base of North America, it initially contacted the base of the mantle lithosphere, at about 100 km depth. The cold slab chilled and strengthened this mantle lithosphere (which originally lay along the coast), forming a "mantle plate" that subsequently moved without internal deformation or thickening (Figure 3). Although the shear stress applied to this strong edge was limited, the cumulative force set the entire mantle lithosphere of North America in motion to the northeast, at various fractions of the velocity of the Farallon plate (up to 12 cm/year). The mantle lithosphere moved without displacing the crust, from which it was kinematically detached by a weak shear zone in the lower crust. The mantle lithosphere was shortened horizontally and thickened vertically by pure-shear strain, and was left beneath the Great Plains and central lowlands.

After the mantle lithosphere was removed, the Farallon plate dragged directly on the crust. It was also chilled and strengthened, but because of its lower creep-softening temperatures, it did not become rigid. Rather, a simple-shear zone in the lower crust accommodated the relative velocity over a huge area. Since the scale thickness of this shear zone was only 2-3 km and the relative displacement was of order 3000 km, shear strains of order 1000 accumulated in the lower crust. A thin boundary layer of crust was transported with the Farallon plate, causing crustal thinning in west, and crustal thickening in the east (Figure 4). This crustal thickening in the Rockies gradually reversed the initial subsidence that had been caused by the weight of the slab.

At the surface, horizontal compressive stress was generated by these basal tractions, as suggested by Dickinson and Snyder (9). This stress was largest at the northeastern margin of the region of flat subduction, as required by the stress equilibrium equation. In addition, this relatively depressed region had less vertical topographic stress to resist shortening. This is why surface shortening strains form an arc from Montana to New Mexico in all models, regardless of rheological parameters. In this particular model, the average net strain in Wyoming has been adjusted to 11% shortening by experimenting with the crustal rheology. The maximum strain rates at the surface occurred when the flat slab area reached maximum width, in the early Paleocene (Figure 2).

Later, the region of flat subduction contracted (this happened by sinking, not reversal, of the Farallon plate). This exposed the base of the crust to upwelling hot asthenosphere over a vast region west of the Rockies (Figure 3). Heating from below decreased the strength of the crust at the same time that isostatic rebound increased its elevation, and therefore the tendency for gravitational spreading. Thus, crustal extension followed several million years after the removal of the flat slab, but only in the region from which the mantle lithosphere had been removed. This extension was simultaneous with flat subduction further west, which was not sufficient to overcome the spreading tendency of the thickened, heated, and uplifted crust.

Comparison to Geologic History

The prediction of vertical crustal movements is the least successful part of the model. The Rocky Mt. foreland region actually: (i) subsided by as much as 3 km sometime during 84-66 m.y.B.P. (10, 16); (ii) rose past sealevel slightly before 65 m.y.B.P., stopping at perhaps 1 km (5); and (iii) acquired its present 2 km elevation after 30 m.y.B.P., perhaps only a few million years ago (5). This model shows a similar pattern, but with offsets in the timing: the crust is (i) depressed to -4.5 km at about 67 m.y.B.P., when the combination of the flat slab and the thickened N. American lithosphere underthrusts it; (ii) uplifted to sealevel (which meanwhile had dropped by 100-200 m (19)) during 66-55 m.y.B.P. by a combination of pure shear, arrival of imported crust, and retreat of the flat slab; and (iii) left at about 1 km when the model ends at 30 m.y.B.P.. If we assume that the overthickened mantle lithosphere was later removed by convection (20) or delamination (21), then the model would be consistent with a second late uplift to 1.5 km or 2.5 km, respectively. Thus, the most important discrepancy is that events are about 10 m.y. late in the model. It is possible that some of this error is inherited from incorrect plate motions and/or positions used in the bottom boundary condition.

The model does very well in predicting the history of surface strain (Figure 5). During 70-65 m.y.B.P. (latest Cretaceous), it predicts a belt of slow shortening running from the Montana/Idaho border south through central Utah and Arizona. In fact, the earliest Laramide structures known are the western ones: the Blacktail-Snowcrest and Madison Ranges in Montana were formed in the latest Cretaceous (22, 23), and thrust-faulting in southern Arizona is also partly Cretaceous (24). The central part of this belt overlaps the pre-existing Overthrust Belt from the waning Sevier Orogeny, which in fact experienced its last major shortening about 70 m.y.B.P. (25).

Then, predicted shortening moves into a long arc in Montana-Wyoming-Colorado-New Mexico and remains there from 64 m.y.B.P. (early Paleocene) through 45 m.y.B.P. (mid-Eocene) with waning intensity. As shown in Figure 5, there is a striking resemblance to the well-known belt of classic Laramide basement uplifts in these states, including coincidence of their long axes (which would be perpendicular to shortening) and their ages (26).

The last phase of compression around 40 m.y.B.P. is displaced SW into northern Utah, and the axis of shortening is rotated counterclockwise to N35E. This seems to be consistent with the late formation of the Uinta Range at an anomalous trend, which was pointed out by Gries (26).

The model predicts no later compression, except in southern Arizona. Meanwhile, crustal extension, which is predicted in British Columbia throughout the model, begins to expand at 50 m.y.B.P. and spreads SE into Washington by 45 m.y.B.P., into Montana/Idaho by 40 m.y.B.P., into western Wyoming by 35 m.y.B.P., and all the way through New Mexico to Chihuahua by 30 m.y.B.P..

This extensional phase is known in geology as the beginning of the Basin-and-Range Orogeny. In the north, it is expressed by formation of "metamorphic core complexes" with low-angle normal faults: in Washington at about 51 m.y.B.P. (27) and along the Montana/Idaho border at about 45-40 m.y.B.P. (28); so the model predictions are matched here. In Wyoming and Colorado, there are less dramatic high-angle normal-faults, but they are of Miocene age (the model is 12 m.y. early). In the south, extension is expressed by formation of the Rio Grande rift, beginning (29) at 32 m.y.B.P. (the model is correct).

The model ends at 30 m.y.B.P., because it is not yet equipped to model the contact of North America with the Pacific plate and formation of the San Andreas fault system. However, the Basin-and-Range orogeny continued and spread up to the present day. In advance of further calculations, it is safe to predict that some degree of extension should affect all areas which lost their mantle lithosphere and were therefore eventually exposed to hot asthenosphere. Figure 3 shows that in fact there is a remarkable agreement between the area swept free of sub-crustal lithosphere and the area subsequently involved in late Tertiary faulting. This suggests that removal of lithosphere by horizontal subduction is the key factor which controls the extent and timing of the Basin-and-Range Orogeny.

Conclusions

The correspondence of model predictions to actual geology is sufficiently close to show that the hypothesis, that horizontal subduction caused the Laramide orogeny, is correct. The Rocky Mountain thrust- and reverse-faults formed in an environment of E-W to NE-SW compressive stress caused by the viscous coupling between the Farallon plate and the base of the North American crust. Nonuniform crustal thickening by simple-shear transport also caused relative uplifts, so this model is consistent with both of the range- forming mechanisms that have been inferred (1). A new proposal arising from this simulation is that horizontal subduction also caused the subsequent extensional Basin-and-Range Orogeny, by stripping away the mantle lithosphere so that the crust was exposed to hot asthenosphere after the oceanic slab dropped away.

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Table 1
Parameters of the North American Plate

	Parameter	Crust	Mantle	Units	Ref.
μ	friction coefficient	0.85	0.85	none	(15)
α	viscosity factor	172	6,390	Pa $s^{1/n}$	
β	(activation energy)/nR	14,000	18,300	K	
n	stress exponent	3.0	3.5	none	(15)
ε	(activation volume)/nR	0.	5.8×10^{-7}	K Pa^{-1}	(15)
σ_p	plasticity stress	5×10^8	5×10^8	Pa	(15)
η_{max}	maximum viscosity	3×10^{24}	3×10^{24}	Pa s	
	initial thickness	3.3×10^4	7×10^4	m	
	radioactive heating	6.1×10^{-7}	1.6×10^{-8}	J $m^{-3}s^{-1}$	
	thermal conductivity	2.5	4.06	J $m^{-1}s^{-1}K^{-1}$	
	thermal diffusivity	1.2×10^{-6}	1.2×10^{-6}	m^2s^{-1}	
	initial heat-flow at top	0.054	0.034	J $m^{-2}s^{-1}$	
	density at STP	2988	3248	kg m^{-3}	

Figure Captions

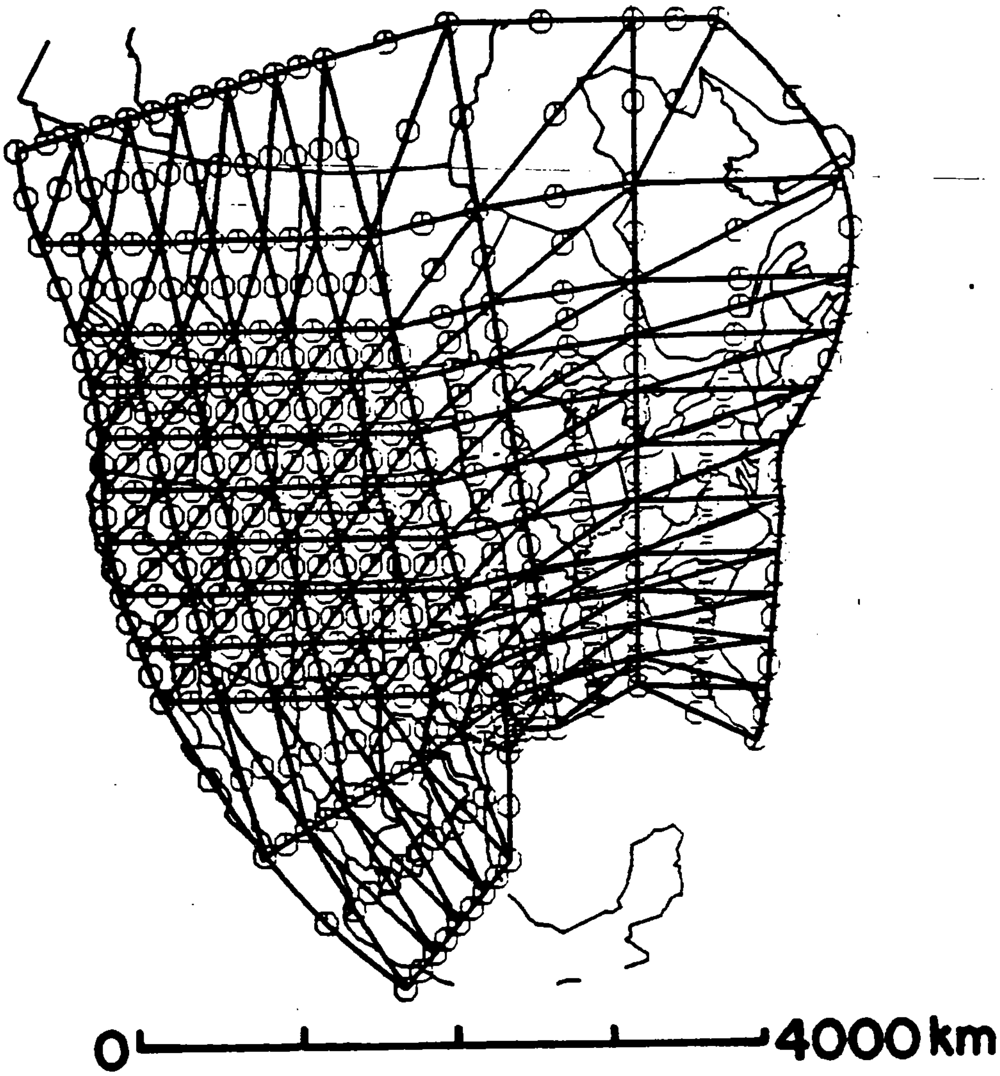
Figure 1. Grid of finite elements used to represent the crust of North America. An identical second grid underneath is used to represent the mantle layer of the lithosphere. Each layer is composed of 198 isoparametric triangles, within which quadratic functions are used to represent velocity and layer thickness. Each triangle contained 7 integration points (not shown) for all area and volume integrals. In this and other figures, the present-day map of states is for general location reference; no attempt is made to restore state lines to former geometries.

Figure 2. Locations of flat subduction through time. The labelled curves separate the area underlain by flat-subducting oceanic slabs (to the west) from areas underlain by normal asthenosphere (to the east). Thus, these lines are the locations at which the oceanic plates separated from North America and descended into the asthenosphere. Curves are labelled in m.y.B.P.. More complete data and maps in reference (4).

Figure 3. Left: final (middle Oligocene) displacement and thickness of the mantle layer of North American lithosphere. Dashed curve shows its former extent, from Figure 1. Thickness is contoured with interval 20 km. Right: summary map of post-Oligocene faults in North America, from Stewart (18). Most (except the San Andreas system) were formed in the extensional Basin- and-Range Orogeny that followed the end of this model. Note the correspondence between the area cleared of mantle lithosphere and the area that was later extended.

Figure 4. Final (middle Oligocene) crustal thickness. Contour interval 5 km. Crust has been transported by simple shear from the coastal region, where it is thinned, to the longitude of the Rockies, where it is thickened by up to 100%. This thickening isostatically supports the present regional elevation (up to 2 km) of the Rockies. Thickening in the Great Plains was actually greater than in this model. Note that the map of state lines is for the present, and that late Tertiary Basin-and-Range extension (not modeled) will bring the coastlines into congruence.

Figure 5. Maps of greatest horizontal principal strain rates and directions at the surface for 4 representative times in the model. Contour interval 10^{-16}/s. Zero contours are dashed. Negative values and converging arrows show compression and imply thrust- faulting perpendicular to arrows; positive values and diverging arrows show extension and imply normal-faulting perpendicular to arrows. Next to the Paleocene (64 m.y.B.P.) map, an inset shows locations of Laramide basement uplifts, in solid black. Note that the map of state lines is for the present, and that late Tertiary Basin-and-Range extension (not modeled) will bring the coastlines into congruence.



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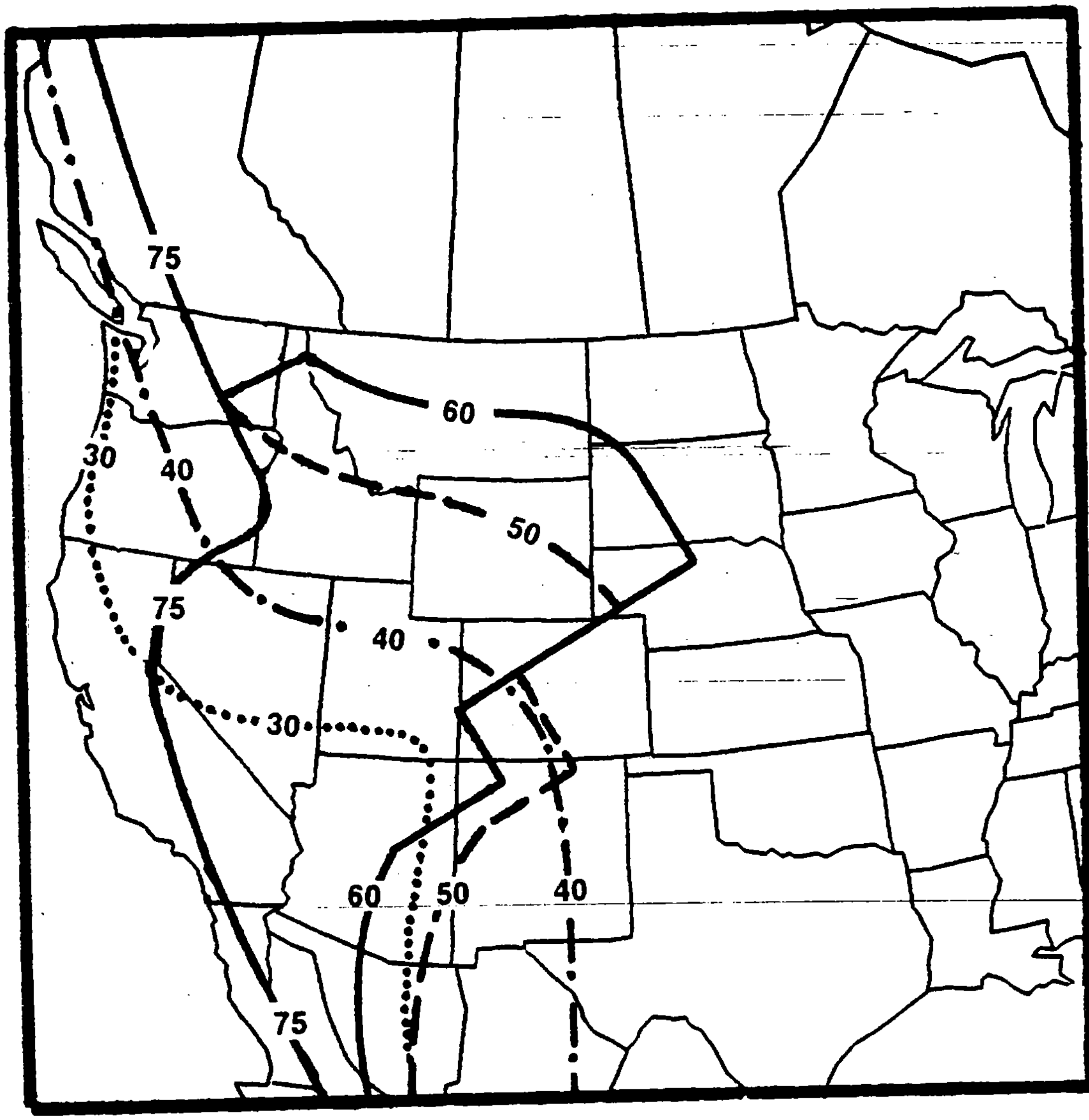
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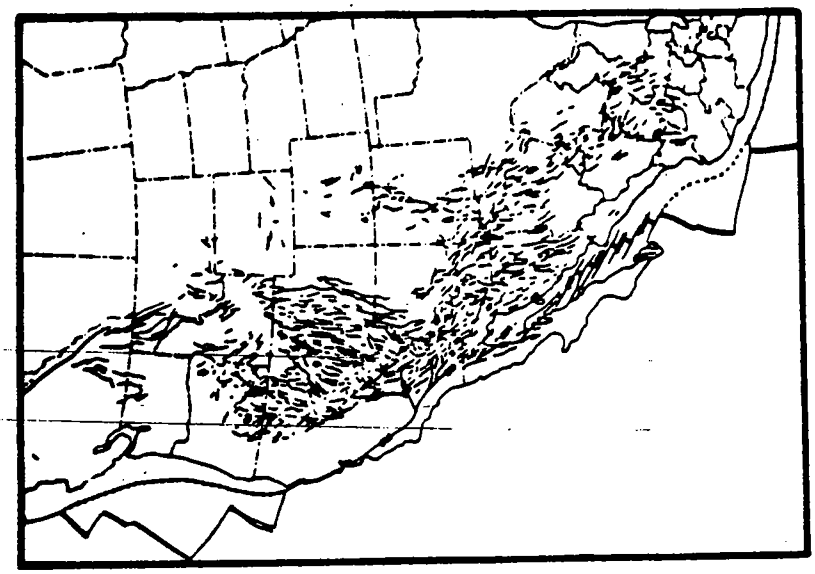

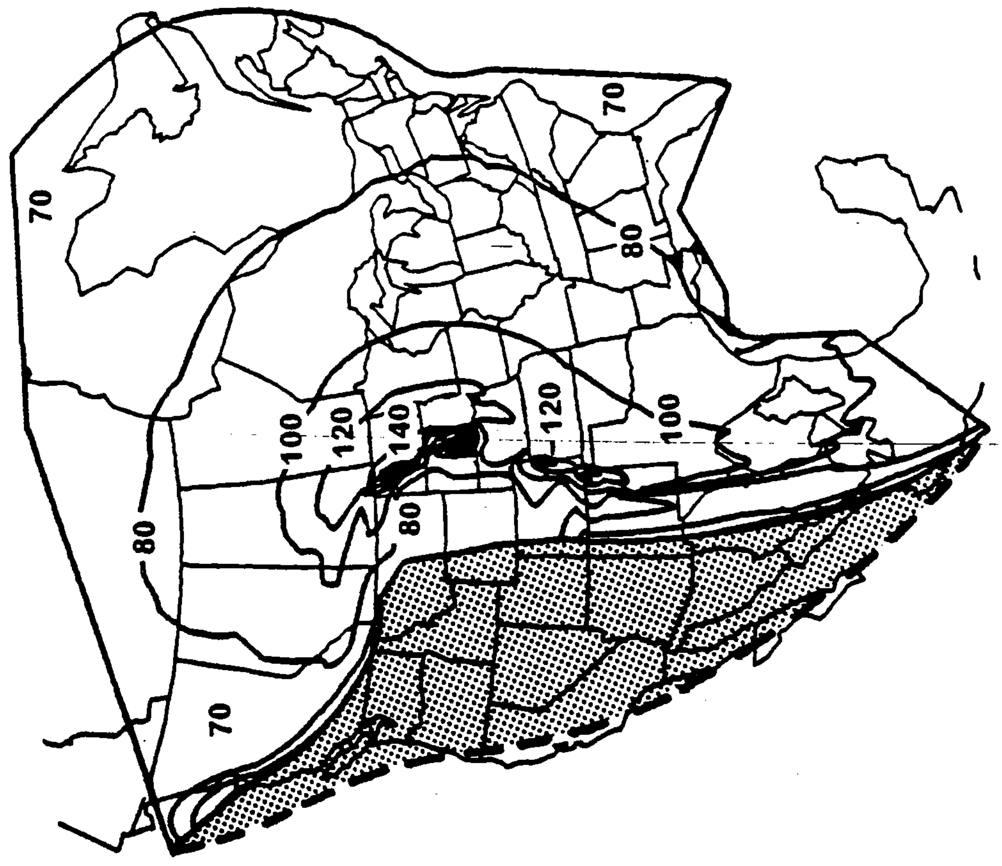
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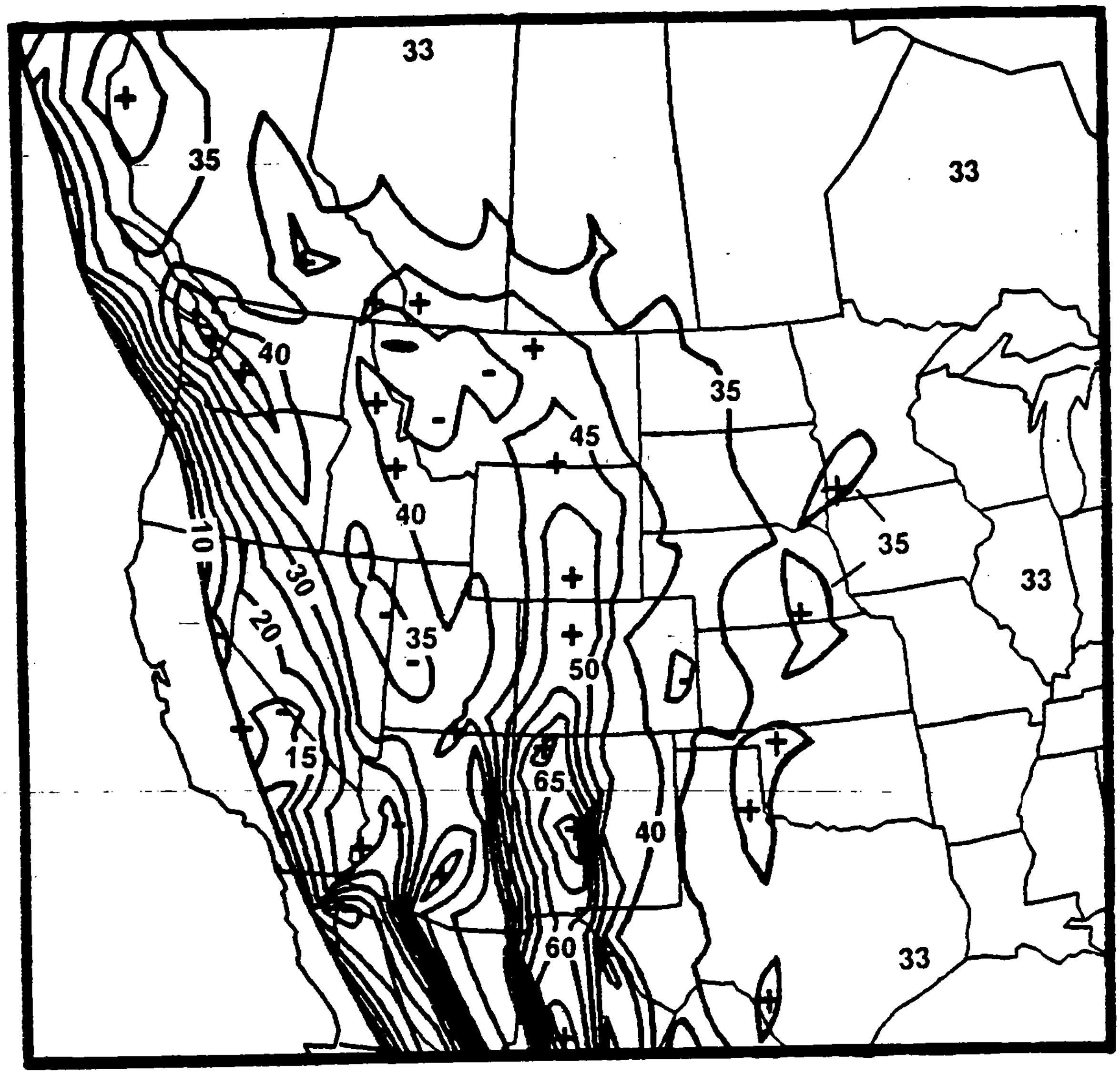
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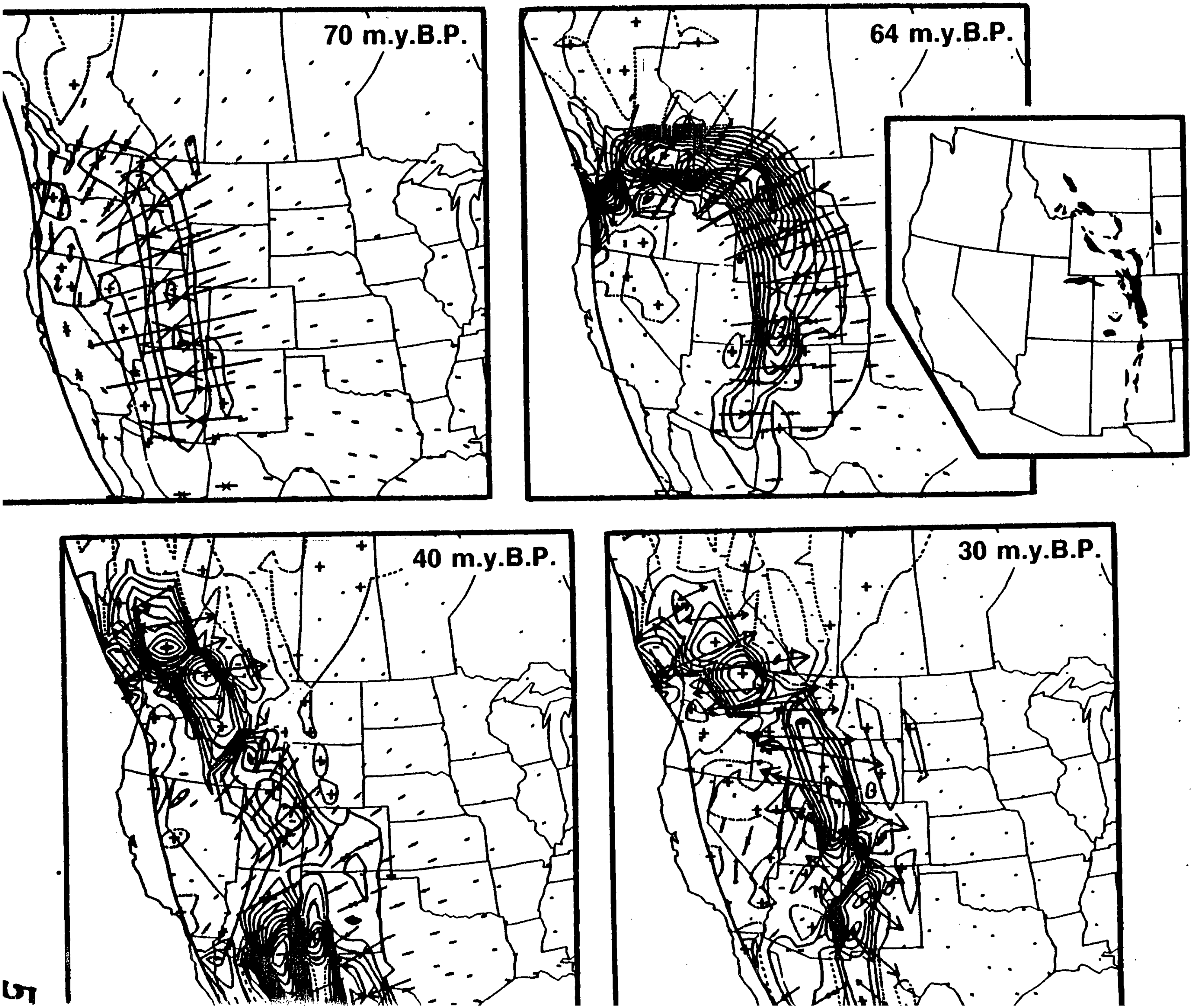

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