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Cause of tectonic reactivation and subtle uplifts in the Rocky Mountain region and its effect on the stratigraphic record

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ABSTRACT

Stratigraphic studies show that widespread, albeit subtle, topographic uplifts have occurred throughout the U.S. Rocky Mountain region during middle and late Mesozoic time. Individual paleotopographic highs generally have wavelengths on the order of tens of kilometres and amplitudes on the scale of metres. In places, these features correspond to sites of previous or later deformation and so suggest the possibility of tectonic reactivation along pre-existing zones of structural weakness in the crust. Observations of modern features and modeling studies indicate that uplifts of small magnitude are both a common and expected result of changes in intraplate stress levels. Finite-element modeling suggests that broad-wavelength, low-amplitude topography can develop in a heterogeneous elastic plate under high-stress magnitudes along preexisting faults and under low-stress magnitudes where zones of high and low strength are juxtaposed. Although these features are small, they have sufficient topography to bring about large changes in paleocurrent patterns for low-gradient streams and significantly affect isopach patterns. Although subtle uplifts from this cause are difficult to predict, they are likely a common occurrence and may be the most dramatic features in an area that is otherwise tectonically quiescent. Thus, subtle uplift may be a background noise rather than a harbinger of regional orogenesis.

INTRODUCTION

A diverse data set—mostly from studies of paleocurrent, sediment-thickness, and facies patterns—suggests that subtle topography developed across the U.S. Rocky Mountain region throughout Phanerozoic time, especially during the Late Jurassic through middle Cretaceous (Fig. 1). For example, detailed study of the Morrison Formation led Peterson (1984, 1986) to suggest that small-scale crustal movements were affecting the western part of the Colorado Plateau in central Utah during Late Jurassic time. These features had wavelengths of several tens of kilometres and apparent amplitudes, determined from isopach patterns, of a few tens of metres. Similar data led others to suggest that generally small-scale topography was developing during Late Jurassic through Early Cretaceous deposition in Montana, Wyoming, Colorado, New Mexico, and adjacent areas (Fig. 1; Merewether and Cobban, 1986; Meyers et al., 1989; Holbrook and Dunbar, 1992; DeCelles, 1986; Schwartz and DeCelles, 1988).

A most comprehensive data set comes from middle Cretaceous formations of the central Rockies (Merewether and Cobban, 1986), where recurrent local deformation is recorded throughout the western interior by

hiatuses determined by biostratigraphic data and facies patterns. The inferred highs were intermittent and, in many cases, follow the trend of established structural highlands. These features are inferred to represent marine topography that developed as a result of regional tectonism associated with partitioning of the subsiding foreland basin adjacent to the Sevier orogenic belt (Merewether and Cobban, 1986). These and other uplifts were generally widely spaced and may not have been active at the same time. Information, where available, indicates that individual features were of broad wavelength (kilometres to tens of kilometres), generally of relatively low amplitude (metres to tens of metres), and range from linear to equant shaped. In most of these cases, differential uplift has been associated with established sites of deformation that occurred previously or where deformation subsequently took place, such as the Laramide orogeny (e.g., Maughan and Perry, 1986; Schwartz and DeCelles, 1988). Although suggestions of subtle tectonics during times otherwise quiescent are not limited to the Rocky Mountain region, they are pervasive in the Jurassic-Cretaceous strata of that region. We suggest that reactivation and uplift along older

Figure 1. General distribution in Rocky Mountain region of inferred uplifts of Late Jurassic (outlined) and Early to middle Cretaceous (solid) age. Location of uplifts from Holbrook and Dunbar (1992), Lorenz (1982), Merewether and Cobban (1986), Peterson (1984, 1986), Schwartz and DeCelles (1988), and references therein. Inference of uplift comes from isopach, paleocurrent, and unconformity patterns.

structures may be facilitated by small changes in intraplate stress fields. As such, subtle tectonics are common and form as the natural consequence of changes in plate-tectonic motions, but offer little insight into regional tectonic history.

INTRAPLATE STRESS AS A CAUSE OF REACTIVATION

Plate-motion processes require the generation of intraplate stress fields that change through time in magnitude and direction (Zoback, 1992). Besides local or regional tectonics, uplift and subsidence can occur because of changes in stress fields caused by far-field processes, such as plate-margin forces, or by local effects, such as intraplate

Figure 2. A: Central part of finite-element mesh crossed by dipping fault defined by slippery nodes along which differential motion (faulting) is allowed. **B:** Surface deflection of elastic plate for three cases with different fault dip (δ). Sharp shape of central developing escarpment is an artifact of mesh and will, in reality, be smoothed through erosional effects. Model parameters are EET = 20 km, Young's modulus = 70 GPa, fault depth = 10 km, and friction coefficient = 0.5.

loading and flexure. Changes in the magnitude and direction of intraplate stresses can amplify previous topography, develop relief along weak zones in the crust, and produce changes in lithospheric rigidity.

Modeling and field studies show that intraplate stress fields are present throughout the interiors of plates. Intraplate shortening associated with collision and deformation in the Appalachians is recorded in cratonic stratigraphic sequences up to 1700 km in front of the fold-and-thrust belt (Craddock et al., 1993). Depending on the geometry of the plate and the distribution of dynamic processes around it, areas with high stress magnitudes may exist. In these areas, the occurrence of large changes in the intraplate stress field may cause extensive deformation of the entire lithosphere on a scale of hundreds of kilometres (Stephenson and Cloetingh, 1991). However, small stress changes may also lead to intraplate deformation, albeit on a lesser scale (Karner, 1986; Cloetingh, 1988).

Releveling surveys suggest that even tectonically tranquil parts of the continents move up and down at relative rates of several millimetres per year (Brown and Reilinger, 1986). Other evidence of intraplate deformation includes seismicity, active surface faulting, and other deformation that may produce several metres of relief (Slemmons, 1991). In some regions, inferred motion is corroborated by geomorphic observations, such as adjustments of river systems (Adams, 1980; Schumm, 1986). Although many causes generate recent vertical motion, such motion has been linked in some cases to temporal variations in the intraplate stress field (Adams, 1980).

NUMERICAL MODELING OF INTRAPLATE DEFORMATION

To determine whether small variations in intraplate stress may affect the stratigraphic record, we carried out finite-element calculations on elastic-plate models, which contained various types of inhomogeneities (i.e., faults, basins, plutons) and which were

subjected to a compressional horizontal load. These models in no way reproduce exactly the deformation of a specific area, but they help us examine the response to intraplate stresses of local weakness zones of several types and scales. Real deformation processes occur in much more complex settings within three-dimensional anisotropic plates that have depth-dependent rheologies and that deform elastically, viscously, and plastically as they are subjected to three-dimensional and time-varying loads at the boundaries as well as in the interiors.

The finite-element models were kept simple: two-dimensional plane-strain cross sections of elastic plates with constant thickness. Inhomogeneities in the upper and lower parts of the plate are represented by blocked areas that have a different Young's modulus than their surroundings. Preexisting faults have a simple, planar geometry with dip, length, and friction that can be varied. Plate models are supported by an elastic foundation representing isostatic restoring forces.

For the computations, we used a modified version of a code developed by Melosh and Williams (1989) suitable for analyzing deformation including the effects of friction, along and over faults. Because it is difficult to know the structural configuration of the Rocky Mountain region prior to Laramide time, we used a generalized Laramide structure for modeling purposes. We constructed a model containing a single fault, which extended from the surface to a depth of 10 km (Fig. 2A). We modeled three cases for fault dips of 30° , 45° , and 60° . For the strength of the plate, we assumed a Young's modulus of 7×10^{10} Pa (70 GPa), which is a typical value for metamorphic rocks, and a friction coefficient of 0.5, which is an average for many kinds of rocks (Carmichael, 1989). The left edge of the plate is fixed horizontally. On the right edge, a constant-velocity boundary condition is imposed, assuming an initial strain rate of internal deformation of 10^{-16} s $^{-1}$. Gravitational stresses are included, assuming an isotropic state of stress.

Shortening of the plate will lead to a compressive stress field with increasing magnitude, which at a critical value will start unlocking the fault. Figure 2B illustrates surface deflection of the plate for the three fault dips after 1.25 m.y. (0.4%) of shortening. In addition to a sharp central escarpment, which in nature would be smoothed by erosion, flexural deflection of the fault blocks develops; however, it develops increasingly less for steeper faults and not at all for the 60° -dipping fault (Fig. 2B). Deflections are on the order of a few tens of metres over a distance of 50 to 100 km, giving topographic gradients of $\geq 10^{-4}$. The stress levels required to produce these topographic gradients suggest that fault reactivation does not start before the effective stress has reached ~ 30 MPa and that fault slip (and accompanying flexural deflection) only reaches significant values for stress levels up to 100–150 MPa. These are large stress levels, implying that, as the model is set up, fault reactivation is not a suitable mechanism for producing small-scale uplifts in tectonically quiescent regions.

We also modeled plates containing zones in which the Young's modulus differs from that of the surrounding plate. These 25-km-wide by 10-km-deep zones of anomalous strength are located in the upper or lower parts of the plate (Fig. 3, A and C). Generally, these zones might correspond to sedimentary basins or plutons within the crust.

We analyzed models with different two-dimensional strength contrasts in the plate: 10 vs. 20 GPa (strength ratio $\gamma = 2$) and 10 vs. 100 GPa ($\gamma = 10$). The lower values are representative for sedimentary rocks, and the higher values are representative for crystalline or plutonic rocks (Carmichael, 1989). In Figure 3, B and D show the calculated surface deflections of the elastic plate for these strength contrasts for the four end-member geometries in A and C.

The surface deflections of the upper and lower configurations of Figure 3, A and C, illustrate the importance of the position of the deviating area within the plate. A change from an upper- to a lower-plate position results in a reversed response of surface deflection, i.e., subsidence vs. uplift. However, the magnitude of deflection is not affected significantly by a change in plate position. A distinction also exists between a weak zone within a strong plate (Fig. 3A) and a strong zone within a weak plate (Fig. 3C), which reverses the response in the direction of surface deflection. A local weak zone produces a surface deflection that is about ten times greater, although only for the case where $\gamma = 10$. Also, changing the strength contrast between the zone and sur-

Figure 3. Surface deflection of heterogeneous elastic plate. Top panels (A and C) show central sections of finite-element mesh of four end-member cases: deviating area in upper vs. lower part of plate and (A) weak area (fine dots) in strong plate (unpatterned) vs. (C) strong area in weak plate. Bottom panels (B and D) show surface deflections of elastic plate after 0.25 m.y. of shortening for overlying models and for each model for two cases of different strength contrasts in Young's moduli: 20 GPa vs. 10 GPa ($\gamma = 2$) and 100 GPa vs. 10 GPa ($\gamma = 10$). "Upper" and "lower" denote curves for weak zones located in upper or lower part, respectively, of plate. Note changes in scale between different parts of figure.

rounding plate affects the amplitude of the surface deflection up to a factor of ~ 5 .

The differences in amplitude are probably caused by localization of the elastic strain in the relatively weaker part of the plate. Figure 3A shows that most of the strain is concentrated in a small area and that the amplitudes of deflection are relatively high. In addition, Figure 3A shows that, when localized, a change in strain produced by a change in the strength contrast will produce a more pronounced change in amplitude. Figure 3C shows that when the strain is distributed over a much larger zone, less deflection occurs.

In all cases, a flexed surface develops with deflections of a few metres over a width of ± 50 km (Fig. 3D) to several tens of metres over a width of ± 100 km (Fig. 3B). These surface deflections correspond to topographic gradients of $\sim 10^{-4}$. For the states of deformation shown in Figure 3, stress levels in the plate are on the order of a few megapascals up to ± 10 MPa. These stress levels are smaller than those expected in tectonically active areas (Zoback, 1992). Thus, topography can develop within nonisotropic plates during times of changing intraplate stress in the absence of local major tectonic events.

DISCUSSION

Changes in intraplate stresses add a small, but stratigraphically significant, component of uplift or subsidence to preexisting topography and/or zones of weakness within the lithosphere. Both the model and modern examples show that relief is small with modest variations in stress (tens to hundreds of bars) but can be substantial when stress levels are high (kilobars). Certainly, vertical motion on the scale of metres to a few tens of metres is reasonable under many geologic conditions. The rate at which topography develops or changes is directly related to the rates at

which in-plane stress vectors change. Stress changes of a few tens of bars are likely to be fairly common whereas changes on the order of kilobars are large in scale and less common.

Model results suggest that differential movement can be focused in zones of preexisting weakness in the crust. The model predicts more relief (up to tens of metres) under more modest stress levels (tens to hundreds of bars) in a heterogeneous plate containing lateral changes in elastic properties than in a homogeneous plate with preexisting faults. The greater the change in elastic properties, the larger the relative movement. Because such lateral changes in rheology may occur across faults, these combined effects may interact to localize uplift in these zones of weakness. Uplift patterns can range from linear to equant shaped depending on the distribution of crustal weakness and flexural effects of local loading.

The specific magnitude and even direction of a stratigraphic response due to intraplate stress variation depend on several variables that influence lithospheric rigidity as well as preexisting topography, isostatic balance, magnitude of the stress field, the number and spacing of faults and their depth, dip angle, and coefficient of friction (Cloetingh, 1988). Our models do not include the many complexities of the real world. Thus, it is difficult to predict and filter out for the past the location, timing, magnitude, and even vertical direction that these features would develop in the rock record. Instead, we have used the model to illustrate the effects of changes in intraplate stresses on uplift. Although it is hard to approximate a model that is reasonable for the Rocky Mountain region, it is worth noting that deep seismic reflection studies (Johnson and Smithson, 1985; Smithson et al., 1978) suggest that Laramide faults have dips of up to $\sim 40^\circ$ and reach

depths of tens of kilometres. These structures are similar to the modeled faults, and, if representative of preexisting crustal structure, the model results may reflect the scale of uplifts in the region during Jurassic through middle Cretaceous time.

Effect on the Stratigraphic Record

The topography variations created in our model have sufficient magnitudes and length scales to significantly affect isopach patterns, facies distributions, and paleocurrents. For example, Heller and Paola (1989) estimated that gradients of rivers that deposited widespread gravel across the western interior during Early Cretaceous time had slopes in the range of 10^{-4} . With such low gradients, it takes very little change in topography to effect a major change in paleocurrent orientation. Of course, with steeper stream gradients, there must be much more significant uplift to force a major change in paleocurrent direction. However, because many aggrading river systems have low gradients and because subtle uplifts of the type described here are expected and common features of regions that are otherwise stable, there may be little tectonic significance to relatively large changes in paleocurrent patterns. The same is true for other stratigraphic indicators of tectonism in the rock record such as shoreline migration, uplift, and erosion of basin margins and changes in isopach patterns on the scale of tens of metres or less of relief over tens of kilometres or more in horizontal distance.

An abundance of evidence exists for subtle uplifts in the Rocky Mountain region during Late Jurassic through middle Cretaceous time. Some of these uplifts overlap the time of shortening in the Sevier fold-and-thrust belt to the west (Fig. 1). Growth of these features may have been accelerated during development of the Sevier belt if thrusting required higher intraplate stress levels.

However, some foreland uplifts likely predate the late Early Cretaceous age of genesis of the Sevier belt (Heller and Paola, 1989). All features predate deformation associated with the Laramide orogeny, which began in Late Cretaceous time (Dickinson et al., 1988), and postdate the arrival of large accreted terranes in western Canada and the adjacent United States (van der Heyden, 1992; McClelland et al., 1992). Sedimentation in the Rocky Mountain region during Late Jurassic to middle Cretaceous time took place on a relatively stable platform or possibly an incipient foreland basin. As a result, small-scale uplifts are the only local deformations that occurred and so dominate the rock record of the region for that period.

CONCLUSIONS

Subtle changes in uplift and subsidence patterns may be precursors to major tectonic events. However, because intraplate deformation is common within a heterogeneous lithosphere, it is best not to attach major tectonic significance to such small-scale features without other corroborating evidence. Because the features generated by intraplate stresses are broad and subtle, they will tend to be best recorded in relatively thin stratigraphic sequences deposited over broad basins that have few other tectonic signals to mask the intraplate effects. In addition, more local differential movement may be concentrated along preexisting zones of weakness in the underlying basement. The western interior basin, which has been a broad region of tectonically quiet and relatively thin deposition for much of the Paleozoic and Mesozoic, is an ideal setting to generate and record such features. Even modest changes in stress levels would produce sufficient topography to strongly affect the stratigraphic record of that time.

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