Beveling the Colorado Plateau: Early Mesozoic Rift-Related Flexure Explains Erosion and Anomalous Deposition in the Southern Cordilleran Foreland Basin

James B. Chapman and Peter G. DeCelles

1Department of Geology and Geophysics, University of Wyoming, Laramie, WY, USA; 2Department of Geosciences, University of Arizona, Tucson, AZ, USA

Abstract Deposition of the Late Jurassic Morrison Formation in a back-bulge depozone and formation of the overlying sub-Cretaceous unconformity above a forebulge mark the birth of the foreland basin system in the central U.S. Cordillera. In the southern U.S. Cordillera, the Morrison Formation is either anomalously thick or absent and the sub-Cretaceous unconformity cuts out progressively older stratigraphy toward the south on the Colorado Plateau. Based on results of 2D and 3D flexural modeling, we suggest that flexural uplift of the northern rift flank of the Bisbee segment of the Borderland Rift Belt can explain these observations. Structural restoration of the sub-Cretaceous unconformity indicates a minimum of 1.5 km of uplift and flexural models with an effective elastic thickness of 55 ± 5 km can reproduce the geometry of the unconformity and rift flank. This implies that effective elastic thickness has decreased between the Jurassic and the present, consistent with hypotheses for uplift and modification of the Colorado Plateau lithosphere during the Late Mesozoic to Cenozoic. Modeling results also predict the presence of a rift-related flexural trough in the Four Corners region of the Colorado Plateau, which may explain above-average thickness of the Morrison Formation. Constructive interference between a flexural back-bulge depozone and a flexural rift-flank trough may help explain anomalously high Late Jurassic subsidence.

Plain Language Summary Continental rifting in southern Arizona during the Jurassic caused the margins of the rift system to be flexed upward and eroded. The erosion resulted in a regionally extensive, low-angle unconformity beneath Cretaceous sedimentary rocks in the southern Colorado Plateau. As the rift margin was flexed upward it caused nearby parts of the rigid crust to flex downward in response, specifically in the Four Corners region of the Colorado Plateau (approximately where the state boundaries of Arizona, New Mexico, Colorado, and Utah meet). The downward flexure in the Four Corners area created a basin where sediments of the Morrison Formation accumulated and are thicker than anywhere else in the western U.S.

1. Introduction

Sedimentation patterns in foreland basin systems provide some of the best constraints on the tectonic histories of Cordilleran-style orogens and are the basis for evaluating a wide range of geodynamic models (e.g., Beaumont, 1981; Dávila et al., 2010; DeCelles & Horton, 2003; Liu & Gurnis, 2010; Miall & Catuneanu, 2019; Painter & Carrapa, 2013). The Mesozoic to early Cenozoic Western Interior Basin is an archetypal retroarc foreland basin system and informs much of the current thinking about Cordilleran systems globally and throughout geologic time (DeCelles, 2004; Jordan, 1981; Kauffman & Caldwell, 1993; Miall & Catuneanu, 2019; Plint et al., 2012). In the central and northern U.S. Cordillera, the onset of deformation in the Cordilleran thrust belt and sedimentation in the eastward adjacent foreland basin system are primarily constrained by deposition of the upper Jurassic Morrison Formation (Figure 1) in what has been interpreted to be the deposits of a back-bulge depozone (DeCelles & Curie, 1996; Fuentes et al., 2009). Supporting this hypothesis is a regionally extensive disconformity that separates the Morrison Formation from overlying Aptian (125–113 Ma) sedimentary rocks and is interpreted to have formed during the passage of a flexural forebulge possibly combined with dynamic uplift (Currie, 1998; DeCelles et al., 2004). Although this interpretation is consistent with stratigraphic data from southern Utah to Alberta (DeCelles, 2004; Miall et al., 2008), it does not explain regional stratigraphic relationships in the southern Colorado Plateau and
regions to the south, where the sub-Cretaceous disconformity becomes a very low-angle angular unconformity that bevels progressively older strata to the south, cutting down-section into lower Paleozoic and Proterozoic rocks in central Arizona and New Mexico. The amount of missing stratigraphic section in this region (ca. 2 km) is incompatible with eustatic changes or the magnitude of erosion associated with uplift of a flexural forebulge (DeCelles & Burden, 1992).

An alternative model to explain the sub-Cretaceous unconformity in the southern Colorado Plateau region is uplift of the northern flank of the Late Jurassic to Early Cretaceous Border (Bisbee) rift belt (Figure 1) (Bilodeau, 1986; Dickinson, 2013; Dickinson et al., 1989). This study investigates whether flexural uplift of the northeast Bisbee rift shoulder (defined as the gently sloping outer flank of the rift highland; Şengör, 2011) is a viable mechanism to explain the geometry of the sub-Cretaceous unconformity and explores potential interference patterns between distal foreland basin-related flexure and rift-related flexure. To model lithospheric flexure, we constructed a pre-Cretaceous paleogeologic (subcrop) map for the Colorado Plateau region (Figure 2) and used the Jurassic-Cretaceous unconformity as a datum or geometric marker. Flexure was modeled analytically in two-dimensions (2D) and numerically in both 2D and three-dimensions (3D) with spatially variable loads and effective elastic thickness.

2. Geologic Background

2.1. Stratigraphy

For the purposes of this study, the pre-Cretaceous stratigraphy of the Colorado Plateau and southern U.S. Cordillera is divided into several units, including the Precambrian basement (predominantly Proterozoic igneous and metamorphic rocks of the Yavapai and Mazatzal lithospheric provinces), Cambrian through Permian sedimentary rocks, the Triassic Moenkopi and Chinle Formations, the Triassic-Jurassic Glen Canyon Group, the middle Jurassic San Rafael Group, and the upper Jurassic Morrison Formation. The middle Jurassic (172–155 Ma) San Rafael Group is located stratigraphically beneath the Morrison Formation and was deposited throughout the southern U.S. Cordilleran foreland (Anderson & Lucas, 1994; Dickinson & Gehrels, 2008). The San Rafael Group crops out in the Colorado Plateau region and correlative rocks, mainly eolian sandstone, are exposed in southern Arizona and northern Sonora, Mexico (González-León et al., 2009; Lawton et al., 2018; Riggs et al., 1993) suggesting the presence of a semi-continuous depositional surface.

In the central U.S. Cordillera, the Morrison Formation is generally <250 m thick, except in the Four Corners region of the Colorado Plateau where it reaches >300 m (Craig et al., 1955; Dam et al., 1990, Figure 1). The Morrison Formation is notable for its symmetric, saucer-like isopach pattern, which contrasts with typical foredeep deposits that thicken exponentially toward the thrust belt (Heller et al., 1986). The symmetrically shaped Morrison lithosome has been interpreted to be a back-bulge depozone within a flexural foreland basin system (Figure 2) and to represent the oldest stratigraphic evidence for the retroarc thrust belt (DeCelles & Burden, 1992; DeCelles & Currie, 1996; Currie, 1998; Fuentes et al., 2009). Sediment provenance and paleocurrent data support this interpretation (Craig et al., 1955; Currie, 1998; DeCelles & Burden, 1992; Dickinson & Gehrels, 2008; Heller et al., 2015; Robinson & McCabe, 1997; Turner & Fishman, 1991). An age-equivalent foredeep depozone related to the Morrison Formation is absent, presumably eroded, and has been called the phantom foredeep for its inferred presence (Long, 2012; Royse, 1993). Radiometric ages of tuffs in the Morrison Formation and magnetostratigraphic studies indicate that it was deposited between ca. 156 and 147 Ma (Kowallis et al., 1998; Maidment & Muxworthy, 2019; Trujillo et al., 2014; Turner & Peterson, 2004). The uppermost part of the Morrison Formation is marked by a several tens of meters thick composite paleosols (Currie, 1998; Demko et al., 2004) and a major Late Jurassic–Early Cretaceous disconformity or paraconformity called the “K unconformity” by Pipirigos and O’Sullivan (1978); this will be
referred to as the sub-Cretaceous disconformity herein. The lacuna represented by the sub-Cretaceous disconformity spans an estimated 20 Myr and has been interpreted to be related to extreme stratigraphic condensation during the eastward migration of the flexural forebulge bounding the west side of the Morrison back-bulge depozone (Currie, 1997, 1998; DeCelles & Currie, 1996).

DeCelles (2004) modeled the shape and position of the Morrison depocenter using the Turcotte and Schubert (2002) analytical solution for 2D flexure of a broken plate (Figure 2), showing that loading of the lithosphere by the Late Jurassic to Early Cretaceous Luning-Fencemaker thrust belt (Oldow, 1984; Wyld, 2002), a westward precursor to the Sevier thrust belt, reproduced the geometry and position of the Morrison lithosome with a flexural rigidity of $10^{24}$ Nm ($\sim 53$ km effective elastic thickness). The flexural modeling predicts $\leq 10$ m of subsidence in the back-bulge region, however, dynamic subsidence in the Cordilleran retroarc region (Gurnis, 1992) could have pulled down the entire load-driven flexural profile, allowing thicker back-bulge deposits to develop (e.g., Catuneanu, 2004; DeCelles, 2012). When combined with an aggradational fluvial gradient extending from the thrust belt across the crest of the forebulge to the craton (Figure 2), $>150$ m of Morrison Formation could have accumulated in the back-bulge region (Currie, 1998; DeCelles, 2004). The flexural modeling also suggested that the forebulge may have had as much as 300 m of positive relief above the regional datum, consistent with the formation of the sub-Cretaceous disconformity above the Morrison Formation.

Overlying the Morrison Formation and the sub-Cretaceous disconformity are Aptian-Albian (ca. 125–100 Ma) sedimentary units, including the Kootenai, Cloverly, Cedar Mountain, Kelvin, Lakota, and Burro Canyon Formations (Currie et al., 2012; Laskowski et al., 2013; Ludvigson et al., 2010; Owen, 1973; Painter et al., 2014; Sprinkel et al., 2012; Tschudy et al., 1984). The Dakota Formation, also known as the Dakota Group or Dakota Sandstone, disconformably overlies the lower Cretaceous units. The unconformity beneath the Dakota Formation is called the sub-Dakota unconformity (e.g., Dickinson, 2013). In most areas, the sub-Dakota unconformity (middle Cretaceous in age) is distinct from the sub-Cretaceous disconformity (early Cretaceous in age; ca. 145–120 Ma; Currie et al., 2008, 2012). However, in southwest Utah, Arizona, and southern New Mexico the lower Cretaceous sedimentary rocks (e.g., Burro Canyon Formation) are absent and the two unconformities are co-located (Figure 3). This study examines the sub-Cretaceous disconformity and the role of flexure during deposition of the Morrison Formation during the Jurassic. The Dakota Formation was deposited in a foredeep depozone and transitions up-section from fluvial/estuarine to shallow-marine strata (e.g., Currie, 1998; Currie et al., 2008; Sprinkel et al., 2012; Ulicny, 1999). In the southern Colorado Plateau region, the Dakota Formation onlaps southward onto the Mogollon Highlands and is time-transgressive, but generally Albian-Cenomanian (ca. 105–95 Ma) in age (Cobban & Hook, 1984; Currie, 1998; Williamson et al., 1993).

### 2.2. Mogollon Highlands and Bisbee Rift System

The Cordilleran foreland basin system was bounded on the south by the Mogollon Highlands, an approximately southeast-northwest trending
paleo-topographic high interpreted to have been the uplifted northern rift shoulder of the Bisbee rift system (Bilodeau, 1986; Dickinson et al., 1989; Dickinson & Lawton, 2001; Lawton et al., 2020). Paleocurrent, provenance, and geochronologic data suggest that the northeast rift shoulder was a major source of sediment to the Morrison and Dakota Formations (Cadigan, 1967; Craig et al., 1955; Dickinson & Gehrels, 2008; Hansley, 1986; Turner-Peterson, 1986; Turner & Fishman, 1991). During deposition of the Morrison Formation, the Mogollon Highlands and Mogollon slope were erosional surfaces incised by north- to northeastward flowing streams and rivers (Aubrey, 1989; Bilodeau, 1986). Sediment provenance data indicate that, at least locally, the Paleozoic sedimentary cover had been stripped off the Mogollon Highlands and that Proterozoic basement was exposed by Late Jurassic time (Dickinson & Gehrels, 2008; Laskowski et al., 2013). During the middle to Late Cretaceous the Mogollon Highlands and slope were onlapped by the Dakota Formation and equivalent rocks. Pre-Dakota or pre-Cretaceous paleogeologic (“subcrop”) maps show that progressively older stratigraphic units are exposed beneath the unconformity closer to the Mogollon Highlands (Dickinson, 2013; Dickinson et al., 1989; this study) (Figure 4). This study investigates potential uplift of the Mogollon Highlands during the deposition of the Morrison Formation during the Jurassic (ca. 155–145 Ma). The Morrison Formation marks the beginning of uplift and erosion of the Mogollon Highlands (Dickinson & Gehrels, 2008; Turner & Fishman, 1991; Turner-Peterson, 1986). Where the Morrison Formation was not deposited, south of ~34°N latitude (present-day coordinates), age-equivalent uplift of the Mogollon Highlands is marked by the sub-Dakota unconformity. Onlap of the Dakota Formation (as young as ~94 Ma) onto the Mogollon highlands, therefore, constrains the duration of this specific uplift event. Renewed uplift of the Mogollon Highlands may have continued after the Jurassic, including during the Laramide orogenic event (ca. 80–40 Ma) (Chapman et al., 2019; Gastil et al., 1992; Karlstrom et al., 2020), but later periods of uplift are not related to the deposition of the Morrison Formation, the sub-Cretaceous disconformity, or the sub-Dakota unconformity.

The 2–3 km thick Bisbee rift basin is part of the larger Borderland rift belt that also includes the McCoy, Chihuahua, and Sabinas basins (Dickinson & Lawton, 2001) (Figure 1). The Borderland rift belt has been interpreted as a back-arc rift related to Farallon slab roll-back and the opening of the Gulf of Mexico (Lawton & McMillan, 1999; Stern & Dickinson, 2010) in a highly oblique (nearly neutral) plate kinematic framework (DeCelles, 2004). The Late Jurassic to middle Cretaceous Bisbee basin overlapped in age with the Morrison Formation (Figure 5). Paleontological data and radiometric dates from the oldest, syn-rift, basal members of the Bisbee Group are 156–150 Ma (Kluth et al., 1982; Olmstead & Young, 2000; Spencer et al., 2011), which demonstrates that Bisbee basin rifting and Morrison deposition were synchronous. Post-extensional thermotectonic subsidence of the Bisbee rift basin commenced during the Early Cretaceous (ca. 130 Ma) and waned during the Albian (ca. 100 Ma) when marine limestone was deposited in the rift basin (Dickinson & Lawton, 2001).

### 3. Methods and Results

#### 3.1. Paleogeologic (Subcrop) Map Patterns

The Late Jurassic topography and geometry of the Mogollon Highlands and Bisbee rift shoulder have been obscured by subsequent processes including Late Cretaceous to Paleogene deformation related to the Laramide orogenic event and Basin-and-Range extension during the Neogene. We use the position of the Bisbee basin as a proxy for the position of the Bisbee rift and use the sub-Dakota unconformity as a proxy for the topographic surface during and immediately after rifting.

First, we constructed a subcrop map beneath the sub-Dakota unconformity for the Colorado Plateau region and southern U.S. Cordillera (Figure 4) using 1:48,000 to 1:250,000 geologic maps. This map is similar to an existing “sub-Dakota” paleogeologic map made by Dickinson (2013), Dickinson et al. 1989 and includes data from parts of the Rio Grande rift, Colorado Plateau transition zone, and the periphery of the Basin-and-Range province. Stratigraphic units were only mapped where the sub-Dakota unconformity is present (i.e., where the Dakota Formation or other lower Cretaceous rocks are not eroded). Stratigraphic units intersecting the subcrop surface are older southward and their contacts strike sub-parallel to the margin of the Bisbee basin, which suggests that the dip of strata was northeastward, perpendicular to the Bisbee basin margin during erosion.
Next, we constructed a cross-section from the Transition Zone in southern Arizona to the Four Corners area of the Colorado Plateau using structural data and unit thicknesses from previously published geologic maps (Figure 6). The cross-section was initially constructed with no vertical exaggeration perpendicular to the trend of the stratigraphic contacts in the subcrop map and has two lateral offsets (A′-to-B and B′-to-C) that were chosen to capture areas with the most data coverage. The cross-section was then restored by removing folding and faulting, chiefly extension related to normal faulting, and then flattening on the sub-Dakota unconformity. Vertical exaggeration was then added to the cross-section to make stratigraphic units and very low-angle structural relationships visible (Figure 6). Assuming that stratigraphic units maintained their thicknesses southward before erosion, the initial restoration suggests ~1.5 km of structural relief (i.e., ~1.5 km of eroded stratigraphic section). Paleocurrent data from fluvial strata in the Dakota Formation in the vicinity of the cross-section (Owen, 1973) demonstrate that the sub-Dakota unconformity surface (base of Dakota Formation or lower Cretaceous rocks) was not flat, but sloped downward to the north and northeast from the Mogollon Highlands. Although the topographic relief and geometry of this fluvial surface during the Late Jurassic are unknown, we estimated it using an ideal equilibrium river profile with standard concavity parameters (Hack, 1975; Whipple & Tucker, 1999) and assumed 1 km topographic relief.
over the ~550 km length of the restored cross-section, consistent with gradients of moderately sized river systems (Sklar & Dietrich, 1998). Our estimate of paleo-topographic relief is conservative, falling in the lower range of global continental rift flank topographic relief (1–4 km; Chéry et al., 1992), and is similar to the amount of relief observed on the flanks of the Rhine graben and Rio Grande rift (Brown & Phillips, 1999; Villemin et al., 1986), which have dimensions similar to those of the Bisbee rift segment. The sum of structural relief and topographic relief suggests ca. 2.5 km of total uplift of the northern Bisbee rift shoulder.

3.2. Flexural Modeling

The asymmetric, concave-up geometry of rift shoulders is a classic flexural isostatic response to unloading (upward-directed vertical force) of the lithosphere and mirrors the flexural response to loading (downward-directed vertical force) of the lithosphere (Stern & ten Brink, 1989; Weissel & Karner, 1989) (Figure 2). In comparison to flexural profiles of foreland basin systems (DeCelles & Giles, 1996), the rift flank is analogous to the foredeep and the flexural trough is analogous to the forebulge. Flexural troughs in rift systems have also been referred to as “outer lows” and “hinterland basins” (ten Brink & Stern, 1992; van Balen et al., 1995). The upward-directed force in rift systems is a combination of positively buoyant forces that result from erosion, normal faulting, thermal expansion of the crust, and thinning of the mantle lithosphere, and negatively buoyant forces that result from sedimentation, crustal thinning, crustal metamorphism, and basaltic underplating/intrusion (Chéry et al., 1992; Mueller et al., 2009; van der Beek et al., 2009; Weissel & Karner, 1989). The magnitudes of the forces resulting from these individual processes during the Late Jurassic is unknown and in the following models, only the total magnitude of the upward-directed force (sum of all sub-forces) is considered. Likewise, we do not consider rift flank flexure/uplift related to processes such as small-scale convection (Buck, 1986; van Wijk, Lawrence, & Driscoll, 2008; van Wijk, Van Hunen, & Goes, 2008) or mantle flow and dynamic topography (Daradich et al., 2003), although these processes could theoretically have contributed to the total load experienced by the elastic portion of the lithosphere.
In addition to the magnitude of the vertical load, the two other main variables that control the geometry of flexural profiles are horizontal forces and flexural rigidity. Forces related to horizontal stresses, sometimes called “tectonic” or “buckling” forces, are often ignored in flexural modeling and when they are considered, it is generally only in contractional systems (e.g., Martinod & Davy, 1992; Tikoff, and Maxson, 2001). We ignore horizontal forces for the following models. We express flexural rigidity as effective elastic thickness using a Young’s modulus of 70 GPa and a Poisson’s ratio of 0.25.

3.2.1. 2D Models

To model flexure in 2D, we used the restored geometry (lower panel in Figure 6) of the top of the Jurassic San Rafael Group (base of the Morrison Formation). The Morrison Formation is not believed to have been deposited south of ca. 34°N latitude whereas rocks (mainly eolian sandstone) correlative with the San Rafael Group crop out locally throughout southern Arizona and northern Sonora, Mexico (González-León et al., 2009; Lawton et al., 2018) suggesting the presence of a semi-continuous depositional surface. We treat the top of the San Rafael Group as a paleo-horizontal surface prior to rifting.

First, we employ the analytical solution of Weissel and Karner (1989) for elastic flexure related to rifting (Figure 7). This model solves the equation for 2D flexure of a plate analytically by treating the vertical force as a line load (approximated using the amount of heave on a planar normal fault) and assuming a constant effective elastic thickness. This class of model is a “broken” lithosphere model in which the second derivative (bending moment) of vertical displacement with respect to distance at the position of the line load is equal to zero. By contrast, the slope (first derivative) is equal to zero at the position of the line load for “continuous” or “intact” lithosphere models. We prescribed a fault dip of 60° (assuming Andersonian normal faulting), an upper mantle density of 3300 kg/m³, a basin fill density of 2600 kg/m³, and we varied the fault heave and effective elastic thickness to match the restored San Rafael Group surface. Acceptable fits to the data (estimated visually) are obtained using an effective elastic thickness of 80 ± 5 km and a fault heave of 2650 ± 50 m (Table 1). The best model produced a flexural trough ~500 km wide, located ~550 km away from the edge of the vertical load, and ~150 m deep (Figure 7). To model the flexure of an “intact” plate, we used the analytical solution for 2D flexure from Turcotte and Schubert (2002). Similar to before, we varied the vertical line load and effective elastic thickness to match the restored San Rafael Group surface. An effective elastic thickness of 50 ± 5 km and an upward-directed line load of 3.5 ± 0.5 × 10¹² N/m resulted in the closest match (visually estimated) between modeled flexural profile and the geometry of the restored San Rafael Group surface (Table 1). The best “intact” model produced a flexural trough ~400 km wide, located ~525 km away from the edge of the vertical load, and ~80 m deep.

Vertical line loads and constant effective elastic thickness are likely not realistic conditions for the flexure of rift flanks. In our next model, we used spatially variable loads and flexural rigidity using a 2D centered finite-difference (numerical) solution from Chapman et al. (2017), which is a “continuous” lithosphere model (Figure 7). Mueller et al. (2009) suggested that two of the most important loads in rift systems are the replacement of crust with mantle during crustal thinning (downward-directed load due to +400 kg/m³ density difference) and the replacement of mantle lithosphere with asthenospheric mantle (upward-directed load due to −100 kg/m³ density difference). Following this approach, we prescribed 10 km of crustal thinning and varied the magnitude of mantle lithosphere replacement/thinning for a 100 km rectangular region beneath the rift, which is the approximate width of the Bisbee basin. We prescribed an effective elastic thickness of 15 km beneath the rift region and varied the effective elastic thickness of the adjacent Colorado Plateau region. The closest match (visually estimated) between modeled flexural profile and the restored San Rafael Group surface was achieved using an effective elastic thickness of 55 ± 5 km and 50 ± 5 km vertical thinning of the mantle lithosphere (Table 1). The best model produced a flexural trough ~250 km wide, located ~525 km away from the edge of the vertical load, and ~65 m deep.

3.2.2. 3D Model

Lastly, we modeled flexure in 3D with spatially variable loads using a second-order finite-difference numerical approximation modified from Cardozo and Jordan (2001) (Figure 8). This model is a “continuous” plate model. The purpose of the 3D model was not to match the restored geometry of the San Rafael Group, but to investigate the interaction of lithospheric flexure in the distal Cordilleran foreland basin with flexure related to the Bisbee rift. We used the results of the 2D models to constrain the 3D model geometry and
parameters. A homogenous effective elastic thickness of 55 km was employed to be consistent with our 2D finite difference modeling and the results of DeCelles (2004) for the Morrison basin. The vertical load related to the Bisbee rift was modeled with a 100 km wide by 750 km long parallelogram located in the geographic position and orientation of the Bisbee basin (Figure 8) with an upward-directed stress of ca. 100 MPa.

**Figure 7.** (Top panel) Results of analytical and numerical 2D flexural modeling. Models were evaluated on their ability to reproduce the restored geometry of the top of the San Rafael Group (base of Morrison Formation) shown in the bottom panel in Figure 6. Modeling parameters and results are also listed in Table 1. Bottom panel) Topographic profiles (solid lines) and fitted flexural models (dashed lines) for rift flanks from a variety of rift systems globally to compare against the results of the 2D flexural model of the top of the San Rafael Group (red lines) presented in the top panel. EET = effective elastic thickness. The Alvarado Ridge is not a rift flank, but an example of a tilted surface produced by epeirogenic uplift that is presented for comparison (Eaton, 2008; Roy et al., 1999). Data sources: Baikal rift (van der Beek, 1994); Red Sea rift (Chen et al., 2015); Transantarctic Mountains (ten Brink et al., 1997), East African Rift, Lake Albert area (Ebinger et al., 1989, 1991), Rio Grande rift, Tularosa Basin area (Peterson & Roy, 2005), Tibet (Masek et al., 1994).

### Table 1

**Results of Flexural Modeling**

<table>
<thead>
<tr>
<th></th>
<th>Best fit effective elastic thickness (km)</th>
<th>Width of flexural trough (km)</th>
<th>Depth of flexural trough (m)</th>
<th>Distance from edge of rift flank to center of flexural trough (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Broken lithosphere, analytical solution (2D)</td>
<td>80 ± 5</td>
<td>500</td>
<td>150</td>
<td>550</td>
</tr>
<tr>
<td>Intact lithosphere, analytical solution (2D)</td>
<td>50 ± 5</td>
<td>400</td>
<td>80</td>
<td>525</td>
</tr>
<tr>
<td>Finite-difference flexural model (2D)</td>
<td>15–55</td>
<td>250</td>
<td>65</td>
<td>525</td>
</tr>
<tr>
<td>Finite-difference flexural model (3D)</td>
<td>55</td>
<td>~475</td>
<td>60–70</td>
<td>~425</td>
</tr>
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</table>
Previous flexural modeling of the Cordilleran foreland basin system during deposition of the Morrison Formation (DeCelles, 2004) used a “broken” plate model, which cannot be easily combined with the 3D “continuous” plate model. We employed two different methods to work around this obstacle, which yielded nearly identical results. For the first method, we simultaneously modeled loading from the Cordilleran thrust belt and Bisbee rift, but moved the position of the thrust belt load westward until the position of the back-bulge depozone matched the location of the Morrison basin. We used the vertical topographic load employed by DeCelles (2004), a rectangle with dimensions 1.5 km high by 160 km wide with a density of 2700 kg/m$^3$, and extended it 1500 km in the third dimension (oriented north-south) to create a block that approximates the geometry of the load related to the Cordilleran thrust belt. The center of the load was positioned ~1000 km west of the center of the Morrison basin. For comparison, the central position of the thrust belt load in the DeCelles (2004) model was located ca. 750 km west of the Morrison basin center.

The second method we used was to take the “broken” plate 2D flexural profile from DeCelles (2004), extend it north-south to make a 3D flexural surface, and then add it to the flexural surface produced by modeling of the Bisbee rift alone. The results of this second method are shown in Figure 8. Results from both methods show constructive interference between subsidence related to the back-bulge depozone and subsidence related to the rift-controlled flexural trough. Maximum subsidence in the back-bulge region is 5–10 m and maximum subsidence in the flexural trough is 60–70 m, and both values are consistent with the 2D modeling. The maximum amount of modeled subsidence is located in the Four Corners region of the Colorado Plateau, coincident with the maximum thickness of the Morrison Formation (Figure 8).

4. Discussion

The two variables we explored in the flexural models are vertical load and effective elastic thickness. Only the effective elastic thickness results are geologically meaningful because of the poor constraints on total uplift/flexure of the rift shoulder as inferred from the geometry of the top of San Rafael Group surface during the Late Jurassic. Although our cross-section restoration (to Late Jurassic time) indicates ca. 1.5 km of structural relief, the topographic relief of this surface is based only on geologically reasonable values of modern river gradients and modern rift flank topography. Our estimates of topographic relief are conservative and thus the vertical load estimates in the best fit models could be considered rough minimum estimates, but we caution against overinterpretation. Furthermore, the ways in which vertical forces were calculated are non-unique and likely not geologically significant. For example, our best-fit 2D model using the Weissel and Karner (1989) method suggests ~2650 m of fault heave, which is similar in magnitude to both the thickness of the Bisbee basin fill (2–3 km) and our estimate for total rift shoulder uplift (2.5 km), but this almost certainly does not represent slip on an actual fault plane or fault system—it is merely the result of a geometric expression used to calculate a total vertical load. Similarly, our estimates of crustal and mantle thinning in the 2D finite-difference model are only used as a guideline to calculate the total vertical load and may not have any geologic significance.

Conversely, we do ascribe geologic importance to our estimates of effective elastic thickness. The wavelength of elastic flexure is independent of the magnitude of the vertical load and fully dependent on flexural rigidity. The effective elastic thickness of our best-fit 2D finite-difference models for flexure of the Bisbee rift flank (55 ± 5 km) overlaps within uncertainty the previous estimate (~53 km) based on flexure of the Cordilleran foreland basin (DeCelles, 2004). The effective elastic thickness of the modern Colorado Plateau
is ca. 40 km in the center of the plateau and decreases to ca. 20 km on the margins (Lowry et al., 2000). If the results of our modeling are accurate, they suggest that the effective elastic thickness has decreased between the Late Jurassic and the present-day. There are multiple possible explanations for the decrease in effective elastic thickness. The continental mantle and portions of the lower crust may have foundered or delaminated from the base of the Colorado Plateau (e.g., Levander et al., 2011). The mantle lithosphere may have been thermally eroded by small-scale mantle convection (e.g., van Wijk et al., 2010) or asthenospheric upwelling (e.g., Karlstrom et al., 2008, 2012; Moucha et al., 2009). The continental mantle and lower crust may have been metasomatically weakened (e.g., hydration) during low-angle subduction (e.g., Feucht et al., 2017; Humphreys et al., 2003; Jones et al., 2015). This study cannot distinguish between these processes, which are all proposed to have occurred during the Cenozoic, but it does help to constrain lithospheric strength in the southern Colorado Plateau region during the Jurassic to Early Cretaceous, prior to tectonic and chemical modification.

### 4.1. Alternative Uplift Mechanisms

Effective elastic thicknesses >30 km are somewhat uncommon in regions of active rifting and are above the global average for continental lithosphere (McKenzie & Fairhead, 1997; Tesauro et al., 2012). However, the flanks of continental rifts that form in old, cold, peri-cratonic to cratonic lithosphere often have effective elastic thicknesses approaching or exceeding 50 km, including the Baikal rift (30–50 km; van der Beek, 1997), the Transantarctic Mountains rift (ca. 80 km; ten Brink et al., 1997), and parts of the East African Rift (45–80 km; Ebinger et al., 1989, 1991). Other mechanisms besides high effective elastic thickness could produce very broad rift shoulders, but they are considered unlikely. Lithospheric necking of ductile lower crust beneath rift shoulders may cause broad uplift (Chéry et al., 1992; Kooi et al., 1992). Ductile thinning of the lower crust in the Colorado Plateau during the Jurassic to Early Cretaceous has not been previously documented or hypothesized. The Colorado Plateau crust and cratonic lithosphere has been interpreted to have been relatively cold and stable during this time and shallow marine sedimentation (e.g., Mancos Shale) suggests that, apart from the Mogollon Highlands, the area was near or below sea level until the Late Cretaceous (Bond, 1976; DeCelles, 2004).

Mantle flow and thermal isostacy related to mantle flow may also cause uplift and regional tilting. For example, a mantle plume has been proposed to contribute to rift flank uplift in the East African Rift near the Afar triple junction (e.g., Pik et al., 2008). However, no independent evidence exists for the presence of a mantle plume or significant mantle upwelling in the Mogollon Highlands area during the Jurassic to Early Cretaceous. Relatively rare, syn-rift basalt sills, dikes, and lava flows are associated with the Bisbee Rift, but the rift system is generally magma poor (Dickinson & Lawton, 2001; Glazner et al., 2008; Lawton & McMillan, 1999; Spencer et al., 2011). Mantle upwelling or dynamic uplift related to mantle flow has been proposed to have uplifted the Colorado Plateau and the Mogollon Highlands during the Laramide Orogeny and during the Neogene (Karlstrom et al., 2008, 2012; Liu & Gurnis, 2010; Moucha et al., 2009; van Wijk et al., 2010). These processes may explain Late Cretaceous to Cenozoic uplift, but cannot explain uplift of the Mogollon Highlands associated with the sub-Dakota unconformity. Likewise, contractional deformation and magmatism associated with the Laramide Orogeny locally thickened the crust (Chapman et al., 2020) and may have uplifted the Mogollon Highlands and produced flexural responses (e.g., Gastil et al., 1992; Karlstrom et al., 2020), but these processes post-date the sub-Dakota unconformity. For example, the transition from rift-related sedimentation (ca. 155–100 Ma) to contraction-related sedimentation (ca. 100–78 Ma) in the McCoy Mountains Formation in southeastern California and western Arizona (Spencer et al., 2011) is younger than the sub-Dakota unconformity and mostly younger than the Dakota Formation that onlaps onto the Mogollon Highlands (Figure 5).

Provenance and paleocurrent data from the Morrison Formation indicate that uplift and erosion of the Mogollon Highlands started during the Late Jurassic (e.g., Dickinson & Gehrels, 2008), around the same time that magmatism in the Jurassic arc (ca. 190–150 Ma) in southeastern California, southern Arizona, and northwestern Mexico was waning (Barth et al., 2008, 2017; Tosdal & Wooden, 2015) (Figures 4 and 5). Cessation of Jurassic arc magmatism and initiation of the Bisbee rift system have been attributed to slab rollback (Bilodeau, 1982; Dickinson & Lawton, 2001; Lawton & McMillan, 1999). Although difficult to test, it is conceivable that mantle upwelling during slab rollback contributed to uplifting the Mogollon
Highlands during the Jurassic. The central U.S. Cordillera experienced surface uplift during the Paleogene, following roll-back of the Farallon slab (Cassel et al., 2018), which may be analogous to the Jurassic arc; however, the uplifted region in the central U.S. Cordillera was much broader than the Mogollon Highlands and uplift was accompanied by widespread magmatism (i.e., the mid-Cenozoic ignimbrite flare up). Widespread magmatism has also been associated with Neogene to present slab rollback in the central Andes (e.g., de Silva & Kay, 2018). Analogous magmatic activity did not occur in the southern U.S. Cordillera during the Late Jurassic to Early Cretaceous. Hence, we favor flexural uplift of the northern Bisbee rift shoulder to explain the sub-Dakota unconformity in the Mogollon Highlands area.

4.2. Deposition in a Flexural Trough

One of the interesting implications of this study is that the southern part of the Morrison Formation may have been deposited in a flexural trough related to the uplifted Bisbee rift shoulder (Figures 2 and 7), which is consistent with paleocurrent and provenance data linking the southern Morrison basin to the Mogollon Highlands. Basins formed in rift-related flexural troughs are uncommon, or at least not widely recognized (e.g., van Balen et al., 1995). Examples include the Kalahari basin associated with the Great Escarpment in southern Africa, the Wilkes basin associated with the west Antarctic rift system, the Manesi, Trichonis, Copais, and Istiea basins associated with the Corinth rift in Greece, and the Estacida basin in the Rio Grande rift (Poulimenos & Doutsos, 1997; Roy et al., 1999; ten Brink & Stern, 1992; van Wijk, Lawrence, & Dri scoll, 2008; van Wijk, Van Hunen, & Goes, 2008; Wanke & Wanke, 2007). The relative scarcity of rift-flank flexural trough basins reported in the literature is somewhat surprising considering the likelihood for preservation, particularly in comparison to back-bulge depozones, another type of flexural basin. Recognition of back-bulge depozones in the stratigraphic record has proliferated in the past two decades although flexural subsidence in a back-bulge region is limited to a few 10s of meters and back-bulge deposits are commonly eroded by the laterally migrating forebulge (DeCelles, 2012). Conversely, subsidence in rift-flank flexural trough basins can reach a few 100 s of meters for common values of rift flank uplift (1–4 km) and effective elastic thickness (10–40 km) (Figure 2). Lower values of effective elastic thickness will produce greater flexural subsidence. For example, Roy et al. (1999) estimated as much as 600 m of flexural subsidence in the Estancia basin in the Rio Grande rift with an effective elastic thickness of

\[ \sim 5 \text{ km}. \]

Rift flanks are also commonly subjected to thermal and isostatic subsidence following rifting that can preserve basin deposits (e.g., Viking graben; Badley et al., 1988; White & McKenzie, 1988). Modeling by Ziegler and Cloetingh (2004) suggests that rift systems are \( \sim 65\% \) thermally equilibrated ca. 60 Myr after the end of rifting, which is similar to the difference in age between the (transgressive) lower Dakota Formation on the northern shoulder of the Bisbee rift and the Morrison Formation (Cobban & Hook, 1984; Turner & Peterson, 2004). In any case, spatial fixity of rifts (in contrast to the extreme lateral mobility of thrust belts and their foreland flexural waves) reduces the erosional potential of forebulge uplift.

The Morrison basin is ca. 75–100 m thicker in the Four Corners region of the Colorado Plateau than elsewhere in the Cordilleran foreland basin (Figure 1), which is similar to the maximum amount of subsidence (60–70 m) predicted by our 2D and 3D flexural models (Table 1). Another way to state this is that ca. 75–100 m of the Morrison Formation (up to \( \sim 50\% \) of the total thickness) in the Four Corners area cannot be explained by subsidence of a back-bulge depozone alone. The flexural models also predict positive interference between a flexural trough and the Cordilleran back-bulge depozone that may have amplified subsidence in this region. Constructive flexural interference (uplift) between a rift and foreland basin has been proposed for the uplifted shoulder of the Rhine graben and the Alpine foreland basin forebulge (Gutscher et al., 1995). The Morrison basin may be another example of flexural interference. The predicted amount of subsidence is less than the actual thickness (up to 325 m) of the Morrison Formation, with the additional thickness interpreted to be related to aggradation up to a stratigraphic base level and possible dynamic subsidence (Currie, 1998; DeCelles & Currie, 1996). In addition to influencing depositional patterns in the Morrison Formation, the intersection of the Cordilleran thrust belt and the uplifted Bisbee rift shoulder affected the paleogeography of the Western Interior Basin throughout the Cretaceous and resulted in a long-lived embayment, or “bight,” in eastern Utah and western Colorado (Van Cappelle et al., 2018).
5. Conclusions

Results of this study indicate that the Mogollon Highlands, interpreted to be the northern uplifted shoulder of the Bisbee segment of the Borderland Rift Belt, experienced 1.5–2.5 km of total uplift during the Late Jurassic to mid-Cretaceous (ca. 145–95), based on assumptions about topographic slope and fluvial gradients. Uplift and erosion of the rift flank during this time are indicated by the regional sub-Dakota unconformity. The Mogollon Highlands may have been uplifted during younger geologic events, like the Laramide Orogeny, but those uplift events are not recorded by the sub-Dakota unconformity. A subcrop map of this unconformity and a cross-section, structurally restored to Late Jurassic time, were constructed and used to model deformation of the northern Bisbee rift shoulder in 2D and 3D using both analytical and numerical methods. The geometry of the rift shoulder can be reproduced by flexural models with geologically reasonable values for vertical loads and flexural rigidity. Although the magnitude of the vertical loads is not well constrained—because the total amount of rift flank uplift is not well constrained—the model estimates for flexural rigidity are geologically significant. Our modeling indicates an effective elastic thickness for the Colorado Plateau region of 55 ± 5 km during the Late Jurassic, within uncertainty of previous estimates based on flexural modeling of the Cordilleran foreland basin (DeCelles, 2004). The results suggest that the flexural rigidity of the lithosphere in the Colorado Plateau has decreased from the Late Jurassic to the present.

In addition to explaining the sub-Dakota unconformity in the southern Colorado Plateau, flexural modeling of the Bisbee rift shoulder predicts subsidence in a flexural trough in the Four Corners region. The best-fit parameters in the 2D and 3D models suggest up to ca. 70 m of subsidence related to rift-flank flexure alone. Constructive interference between the flexural trough and the flexural back-bulge depozone in the foreland basin system may have amplified subsidence in this area. Increased flexure-related subsidence may help to explain the anomalous thickness of the Morrison Formation in the Four Corners area, which is 75–100 m thicker than elsewhere in the Cordilleran foreland basin. The study highlights a rare example of a flexural trough basin and an even rarer example of flexural interference between competing tectonic elements in an orogenic system. It also emphasizes the importance of considering along-strike changes in the tectonic configuration of Cordilleran systems and how they can influence foreland basin depositional patterns.

Data Availability Statement

Data analyzed in this study include isopach maps of the Morrison Formation, available through Craig et al. (1955) and Dam et al. (1990). The subcrop paleogeologic map of the sub-Cretaceous unconformity was constructed using data available through the USGS National Geologic Map Database (https://ngmdb.usgs.gov/ngmdb/ngmdb_home.html), which includes geologic maps published by the USGS, Nevada Bureau of Mines and Geology, Utah Geological Survey, Colorado Geological Survey, New Mexico Bureau of Geology and Mineral Resources, and Arizona Geological Survey.

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