Flexural versus dynamic processes of subsidence in the North American Cordillera foreland basin

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[1] Dynamic subsidence related to subduction is an important process in retroarc foreland basin systems. Dynamic subsidence in the North American retroarc foreland has been proposed as dominant in the Late Cretaceous; however, questions remain about the nature of the subcrustal load and the basin response to such processes. We present new isopach data using 130 data points covering a large portion of the U.S. and a revised flexural analysis of the Sevier foreland. Higher rigidity associated with the Wyoming craton can best model flexural subsidence directly prior to ~81 Ma. We constrain the transition from flexural to nonflexural subsidence to ~81 Ma and correlate this with large Late Cretaceous progradation. The locus of subsidence and the location of the subducted Shatsky oceanic plateau suggest that dynamic loading from viscous flow above the subducting plate is the mechanism driving dynamic subsidence in the retroarc foreland basin of the North American Cordillera. Citation: Painter, C. S., and B. Carrapa (2013), Flexural versus dynamic processes of subsidence in the North American Cordillera foreland basin, Geophys. Res. Lett., 40, doi:10.1002/grl.50831.

1. Introduction

[2] The North American Cordillera extends from southern Nevada, northeast through Utah and then north-northwest through western Wyoming, eastern Idaho, and western Montana, and through the Canadian Rockies, forming together with the South American Cordillera the archetypical retroarc orogenic and foreland basin system (Figure 1). Thrust belt deformation initiated in the western U.S. interior in the Middle to Late Jurassic and continued until the Paleocene [DeCelles and Burden, 1992; Hudec, 1992; Lawton et al., 1993; Currie, 1998; DeCelles, 2004]. Subsidence mechanisms for Cordilleran-type orogenic systems include flexural loading and dynamic processes [Jordan, 1981; Mitrovica et al., 1989; Pang and Nummedal, 1995; Liu and Nummedal, 2004; Liu et al., 2010].

[3] Roberts and Kirschbaum [1995] produced six large-scale foreland basin isopach maps in the western U.S. for the time periods spanning the Late Cretaceous, from the Cenomanian (~98.5 Ma) to the Campanian (~72 Ma). Those isopach maps document a transition from a narrow flexural retroarc foreland basin to a longer-wavelength foreland basin (during the Coniacian to Santonian time, ~88.7–83.5 Ma) and then to a basin with a depocenter more than 400 km east of the thrust front during Campanian time [Roberts and Kirschbaum, 1995].

[4] There is evidence that some dynamic subsidence affected the Western Interior throughout the Cretaceous; for example, Bond [1976] asserted that a eustatic sea level rise was not sufficient to explain the thickness of the Cretaceous marine deposits in the Western Interior, and Mitrovica et al. [1989] state that dynamic subsidence, related to the shallowly subducting Farallon Plate during the Cretaceous, caused the transgression of the Western Interior Seaway. Past flexural modeling and subsidence analyses have been able to closely match the foreland basin profile caused by the tectonic loading of the North American Cordillera during the Early Cretaceous [Jordan, 1981; Heller et al., 1986; Currie, 1998, 2002; Pang and Nummedal, 1995]. The wavelength of the foreland basin profile during the Late Cretaceous lengthens; efforts to explain the cause of the lengthening of the flexural wavelength in the Late Cretaceous and the later migration of the depocenter to the east during the Campanian have produced a variety of hypotheses. Jordan [1981] reported that the early wavelength broadening was caused by the erosion and redistribution of the thrust load, whereas Pang and Nummedal [1995] proposed that dynamic subsidence was responsible for the “middle Cretaceous” shift in subsidence style. Understanding whether this lengthening of the foreland basin wavelength was caused by flexural processes or dynamic subsidence will allow us to better test current foreland basin and geodynamic models.

[5] The Campanian shift in subsidence has been attributed to both isostatic and dynamic processes. Cross and Pilger [1978] proposed that a shallowly subducting plate coupled with the overlying continental lithosphere and caused broad isostatic subsidence in the Sevier foreland basin. Later, researchers attribute the Campanian depocenter shift to dynamic subsidence caused by the cold oceanic lithosphere being subducted at a shallow angle below the overriding plate [Mitrovica et al., 1989; Gurnis, 1993; Liu and Nummedal, 2004] (Figure 2). Recent advances in tomography and inverse modeling have allowed researchers to model dynamic effects on subsidence during the Cretaceous based on the current location of the Farallon Plate, plate motion rates, and mantle viscosities [Spasojevic et al., 2009].

[6] The goal of this study is to more precisely constrain the transition from a narrow flexural foreland basin to a longer-wavelength foreland basin to a nonflexural basin in order to better understand the mechanism underlying dynamic subsidence in the western U.S., test existing models, and provide more detailed basin geometries to constrain geodynamic
models. To do this, we construct four isopach maps dividing Turonian (~93 Ma) to Campanian (~76 Ma) time (Figure 3 and supporting information). Also, we flexurally model the early, Santonian to Campanian, long-wavelength flexural foreland basin and compare the nonflexural subsided basin of the later Campanian and discuss it in relation to the various dynamic topography models that have been proposed [Cross and Pilger, 1978; Mitrovica et al., 1989; Gurnis, 1993; Liu et al., 2010].

2. Isopach Maps

[7] New isopach maps are here constructed using ~130 geophysical well logs, depending on the horizon that was correlated. These isopachs span basins in Wyoming, Utah, Colorado, and New Mexico. The main selection criteria for the well logs are to at least penetrate the top of the Dakota Sandstone, which is ~93 Myr [Molenaar et al., 2002] and to have continuous gamma ray logs from the top of the Dakota Sandstone into Campanian sandstones (~76 Ma).

[8] The thicknesses between horizons are calculated based on correlations (criteria described in Text S1 in supporting information). A well list data file that includes the measured depths to the correlated horizons and the calculated thicknesses is included in Text S2 in supporting information. XY grids were then produced in PETRA® (a well log management and interpretation software). Subsequently, the grids were imported into ArcGIS where the grids were contoured using a nearest-neighbor interpolation algorithm. Finally, contours were smoothed by hand, while still honoring the point data, producing isopach maps spanning ~93–89 Ma, ~89–84 Ma, ~84–81 Ma, and ~81–76 Ma, the total span being within the Upper Cretaceous from lower Turonian to middle Campanian. In order to have a better spatial coverage, these isopachs are extended to the north by using existing isopachs from the literature [Weimer, 1960; Weimer and Flexer, 1985; Roberts and Kirschbaum, 1995]. Isopachs from Weimer and Flexer [1985] and Roberts and Kirschbaum [1995] were used to both guide the trend of the contour lines and the thicknesses present to the north. Weimer [1960] does not include isopach data, but it does include a stratigraphic regional transect; thickness from this transect was used to constrain isopach thickness to the north. The data coverage in the northern part of the isopach maps can be found in Figure S1 in supporting information. The ~93–89 Ma and ~89–84 Ma isopach maps show an eastward

Figure 1. Regional index map of the western U.S. including the Sevier fold-thrust belt, Laramide province, Cordilleran magmatic arc, and major basins. Basins shown are the San Juan (SJB), Paradox (PB), Uinta-Piceance (U-PB), Denver-Julesburg (D-JB), Green River (GRB), Wind River Basin (WRB), Big Horn (BHB), and Powder River (PRB) (modified from DeCelles [2004]).

Figure 2. Cartoon of two models that have been proposed as the cause of the nonflexural subsidence observed during the Cretaceous in the Western Interior. (a) Cross and Pilger [1978] state that subcrustal loading from by a flat subducting plate causes broad wavelength subduction. (b) Dynamic effects of a shallowly subducting plate on the mantle above has also been asserted as the cause of broad wavelength subduction [Mitrovica et al., 1989; Spasojevic et al., 2009; Liu et al., 2010].
migrating, narrow depocenter in the west that shallows to the east (Figures 3a and 3b). The \( \sim 84 - 81 \) Ma isopach map shows a broad depocenter in southern Wyoming that thins to the south, southeast, and east (Figure 3c). At \( \sim 81 - 76 \) Ma, the location of the depocenter shifts \( \sim 480 \) km to the east where it trends north-south in what is now eastern Wyoming and central Colorado (Figure 3d).

[9] Foredeep to back-bulge foreland basin strata from \( \sim 93 - 89 \) Ma are dominated by marine shale deposits with intermittent marginal marine and fluvial deposits. At \( \sim 81 \) Ma, the foredeep strata become coarser, with marginal marine and fluvial deposits becoming more prevalent [e.g., Franczyk et al., 1992].

3. Flexural Modeling

[10] As shown by previous researchers, the observed foreland basin system can be reproduced with flexural calculations from the Late Jurassic to early on in the Late Cretaceous [Jordan, 1981; Yingling and Heller, 1992; Pang and Nummedal, 1995; Currie, 1997, 2002]. However, as documented in this study and in previously published isopach maps [Roberts and Kirschbaum, 1995], at \( \sim 84 \) Ma, the narrow foredeep, parallel to the thrust front, is no longer present. Instead, there is a broad depocenter in northeast Utah, southwest Wyoming, and northwest Colorado, which thins to the south-southwest (Figure 3c). This change in subsidence profile has been attributed to early onset of dynamic subsidence [Gurnis, 1993; Pang and Nummedal, 1995; DeCelles, 2004]. Yet, even though the character of the depocenter has changed, we show that the subsidence profile can still be approximated using reasonable flexural parameters (Figure 4).

[11] The \( \sim 84 - 81 \) Ma isopach map pattern has an \( \sim 1400 \) m thick and \( \sim 300 \) km wide depocenter (Figure 3c). Here the lithosphere is treated as an infinite, unbroken elastic plate with a rectangular load [Turcotte and Schubert, 1982; Angevine et al., 1990]. A variety of parameters were used in trying to most closely match the observed basin profile, and the following parameters produce the closest match. Typically, load dimensions of 400 km wide and 2 km high are used to approximate the Cordilleran thrust belt [Jordan, 1981; Currie, 2002]. The period of \( 84 - 81 \) Ma spans a time after the Wasatch Culmination had been formed and was still building, indicating that a higher tectonic load can be assumed [DeCelles, 1994]. The dimensions used in these

Figure 3. Isopach maps that span \( \sim 93 - 76 \) Ma. Black dots represent well log control. Where there is no well log control, we use previously published isopach maps as reference [Weimer, 1960; Weimer and Flexer, 1985; Roberts and Kirschbaum, 1995]. Maps spanning (a) \( \sim 93 - 89 \) Ma, (b) \( \sim 89 - 84 \) Ma, (c) \( \sim 84 - 81 \) Ma, and (d) \( \sim 81 - 76 \) Ma. Note the lengthening of the basin wavelength in Figure 3c. A– A′ is the location of the profile in Figure 4. Figure 3d includes the proposed location of the Shatsky at 80 Ma [Liu et al., 2010].
calculations are a 400 km wide and 2.5 km high tectonic load. Its density is 2500 kg/m³; the basin fill and mantle density are 2300 kg/m³ and 3300 kg/m³, respectively. All of these parameters are similar to those used in previous flexural studies [Jordan, 1981; Yingling and Heller, 1992; Currie, 2002]. Because of the broad wavelength of the depocenter, here, two high flexural rigidities are used: $3 \times 10^{24}$ N m and $4.1 \times 10^{24}$ N m. B–B’ in Figure 3c is also modeled, showing that the tectonic load height decreases (2 km) to the north and the lithosphere becomes less rigid ($2 \times 10^{24}$ N m).

5. Discussion and Conclusions

[13] This study constrains the transition from flexural subsidence to dynamic subsidence that was previously placed at ~83.5 Ma [Roberts and Kirschbaum, 1995; DeCelles, 2004] at ~81 Ma. Our data show that the shape of the dynamically subsiding basin is not east-west trending, as shown in Roberts and Kirschbaum [1995], but rather north-south trending, similar to what is proposed by Weimer [1970]. Stratigraphically, this transition from flexural subsidence to dynamic subsidence takes place at the start of a regional and widespread progradation of coarse, foreland basin deposits, becoming more incisional higher in the section. A basin subsidence transition from higher accommodation in the west and less accommodation in the east (as is the case in a flexural basin) to higher rates of subsidence in the east and less in the west (as is the case when this dynamic subsidence begins) would produce a rapid progradation as observed in the Campanian [Mitrovica et al., 1989; Pang and Nummedal, 1995; DeCelles, 2004].

[14] Flexure is driving subsidence from ~93 to 81 Ma. However, flexural foreland basin depozones of foredeep, forebulge, and back-bulge [DeCelles and Giles, 1996] should not be applied to the western U.S. for times after ~81 Ma because the foreland basin after this time is controlled by dynamic processes.

[15] Previous researchers have observed the broadening of the flexural wavelength in the Coniacian to Santonian (~88.7–83.5 Ma) and have attributed this to the erosion and redistribution of the tectonic load [Jordan, 1981] and to early dynamic influence [Pang and Nummedal, 1995]. Based on our new isopach maps, this broadening of the flexural foreland takes place at ~84 Ma. Neither the erosion nor redistribution of the tectonic load nor dynamic influence is necessary to explain this wavelength change. Instead, the wavelength change can be explained by an increase in rigidity of the loaded lithosphere. Jordan [1981] assumed a constant flexural rigidity of the lithosphere and approximated it at $10^{23}$ N m. The resulting models fit the observed flexural profile in the Early Cretaceous. Pang and Nummedal [1995] recognized the fact that flexural rigidity likely changes throughout the western U.S. Lowry and Smith [1994] used coherence analysis of gravity and topography to calculate the flexural rigidity of several provinces in the western U.S. Their calculated value for the flexural rigidity of the Archean Wyoming craton was $4.1 \times 10^{24}$ N m, significantly higher than the $10^{23}$ N m used for modeling central Utah. DeCelles [1994] reports that at ~84–75 Ma, the Absaroka thrust breaks out to the east, advancing the thrust belt eastward. This would effectively load the western margin of the Archean Wyoming craton. This also coincides with the broadening of the flexural wavelength as seen in the isopach map (Figure 3c). Furthermore, by 84 Ma, the development of the Wasatch Culmination effectively increased the height of the topographic load. Therefore, a flexural rigidity of $4.1 \times 10^{24}$ N m and tectonic load dimensions of 400 km compared to the hypothesized location of the Shatsky Plateau at 80 Ma (Figure 3d). The isopach map shows that the dynamically subsiding basin overlies the area where the northern portion of the Shatsky Plateau was plunging into the asthenosphere, with a similar north-south trend (Figure 3d).
wide and 2.5 km high are reasonable parameters to use for 84–81 Ma. These parameters match the flexural profile observed in southern Wyoming during that time (Figure 4). To the north, the load height and the plate rigidity decrease to 2 km and $2 \times 10^{11}$ N m, respectively, and closely match the observed profile (Figure 4).

[16] The shallow subduction angle of the Farallon Plate and its effects on the mantle have been asserted as the cause of the dynamic subsidence at ~81–76 Ma [Pang and Nummedal, 1995]. Another hypothesis is that a flat-slab subducting plate can displace the hot asthenosphere, couple with the overlying continental lithosphere, and cause isostatic subsidence [Cross and Pilger, 1978] (Figure 2).

Whereas an oceanic plateau (such as the proposed Shatsky plateau) could possibly cause subcrustal loading similar to what Cross and Pilger [1978] proposed, its location at 81–76 Ma occurs in the location where the hypothesized Shatsky Plateau is plunging into the asthenosphere (Figure 3d), suggesting that dynamic loading from viscous flow above the subducting plate is the mechanism driving subsidence in the western U.S. at this time (Figure 2).

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