Geological Society of America Bulletin

Post-Paleozoic alluvial gravel transport as evidence of continental tilting in the U.S. Cordillera

Paul L. Heller, Kenneth Dueker and Margaret E. McMillan

Geological Society of America Bulletin 2003;115;1122-1132
doi: 10.1130/B25219.1

Notes

Post-Paleozoic alluvial gravel transport as evidence of continental tilting in the U.S. Cordillera

Paul L. Heller[†]
Kenneth Dueker
Margaret E. McMillan[‡]
Department of Geology and Geophysics, University of Wyoming, Laramie, Wyoming 82071, USA

ABSTRACT

The western United States contains three thin but remarkably widespread alluvial conglomeratic units that record episodes of large-scale tilting across the U.S. Cordilleran orogen in post-Paleozoic time. These units are: (1) the Shinarump Conglomerate of Late Triassic age exposed in northern Arizona and adjacent parts of Utah, Nevada, and New Mexico; (2) Lower Cretaceous gravel deposits that overlie the Morrison Formation throughout the Rocky Mountain region; and (3) the gravel-rich parts of the Miocene-Pliocene age Ogallala Group in western Nebraska and adjacent southeastern Wyoming. Paleoslopes of the rivers depositing these units were in the range of 10^{-4} to 10^{-3}, based on paleohydraulic calculations. However, depositional thickness trends of these units are not sufficient to have generated such steep paleoslopes. Thus, long wavelength tilting of the Earth's surface must have occurred for these gravels to be transported. Although these units were deposited adjacent to large tectonic features, including an evolving and migrating continental arc, and, for the Ogallala Group, the northward-propagating Rio Grande Rift, the tilting occurred over wavelengths too broad to be directly generated by these features. These widespread gravel units attest to the interplay between the creation of subduction-related isostatic and dynamic topography and continental sedimentation. Hence, paleotopography, as determined from calculated transport gradients of sedimentary deposits, provides a means of relating constructional landforms to mantle-driven processes.

Keywords: paleohydraulics, tectonics, dynamic topography, U.S. Cordillera, conglomerate, paleotopography, syntectonic sedimentation.

INTRODUCTION

A common characteristic of synorogenic alluvial conglomerates is the relatively limited distance they prograde from their uplifted source areas out into the adjacent basins—most are found within a few tens of kilometers of their associated mountain fronts (Fig. 1, A–G). The restricted distance of gravel deposition reflects the long-term balance between the rate of delivery of coarse sediment supply and basin subsidence rate that acts to trap the gravel. Since on an elastic plate both supply and subsidence are proportional to the size of adjacent mountain belts, it is no surprise that basin sedimentation patterns have a similar length scale.

In contrast, a few conglomeratic units seen in the U.S. Cordillera are unusually widespread in distribution (Stokes, 1950; Stewart et al., 1972a; Heller and Paola, 1989), being deposited over many hundreds of kilometers downstream from their source areas (Fig. 1), yet being relatively thin over their entire extent. These gravels are related to the temporal and spatial pattern of subduction and late Cenozoic uplift of the central and southern Rocky Mountains. We argue that the occurrence of these thin, widespread gravel units does not reflect climatic controls, but must record times of large-wavelength/low-amplitude tilting of the continental interior. Thus, we believe they are the stratigraphic records of deep processes affecting the western U.S. during post-Paleozoic time.

THREE WIDESPREAD CONGLOMERATES

The three widespread gravel units of the U.S. Cordillera are the Shinarump Conglomerate Member of the Chinle Formation (Late Triassic) found in the southwestern USA (Fig. 2A); the Lower Cretaceous conglomerates found over much of the U.S. Rocky Mountains (Fig. 2B); and gravelly parts of the Ogallala Group (Miocene–Pliocene) exposed along the western Great Plains (Fig. 2C).[1]

Shinarump Conglomerate

The Shinarump Conglomerate lies at the base of the Chinle Formation, a fluvial and lacustrine unit of Late Triassic age (Carnian; Lucas, 1993) exposed in the southwestern USA (Stewart et al., 1972a). Rivers of the Chinle Formation flowed generally north from the ''Mogollon highlands,'' an interpreted regional uplands in southern Arizona and adjacent regions (Stewart et al., 1972a; Blakey and Gubitosa, 1983), eventually joining major tributaries from farther east. Together they flowed northwest to the contemporaneous shoreline (Riggs et al., 1996). While the Shinarump Conglomerate is dominantly sandstone, conglomerate is common. Unit thickness varies, but it is typically a few meters thick. Gravel composition includes abundant quartz, quartzite, and chert grains (Blakey and Gubitosa, 1983). The Shinarump Conglomerate forms an irregular sheet that rests erosionally upon older Triassic and Permian units (Blakey and Gubitosa, 1983) and is overlain by bentonitic shales and mudstones of the Chinle Formation (Stewart et al., 1972a). Lo-

[†]E-mail: heller@uwyo.edu.
[‡]Present address: University of Arkansas at Little Rock, Department of Earth Sciences, Little Rock, Arkansas 72204, USA.

[1]Color photos and figures to accompany this paper can be found at: http://faculty.gg.uwyo.edu/heller/pubs.htm.

GSA Bulletin; September 2003; v. 115; no. 9; p. 1122–1132; 8 figures; Data Repository item 2003121.

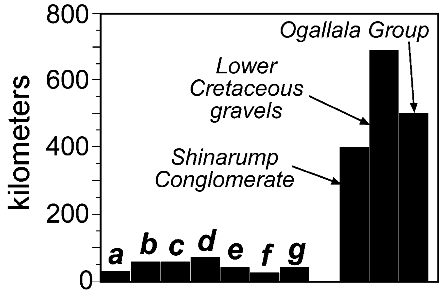

Figure 1. Reported distances of gravel deposition (measured normal to orogenic front) with respect to their orogenic source areas. Examples A–G are from ancient foreland basins (Heller and Paola, 1989); these are, in order, the Alps, Andes, Apennines, Himalayas, Pyrenees, Catalan Coast Range, and Sevier belt. The three unusually widespread alluvial gravels in the U.S. Cordillera are also shown.

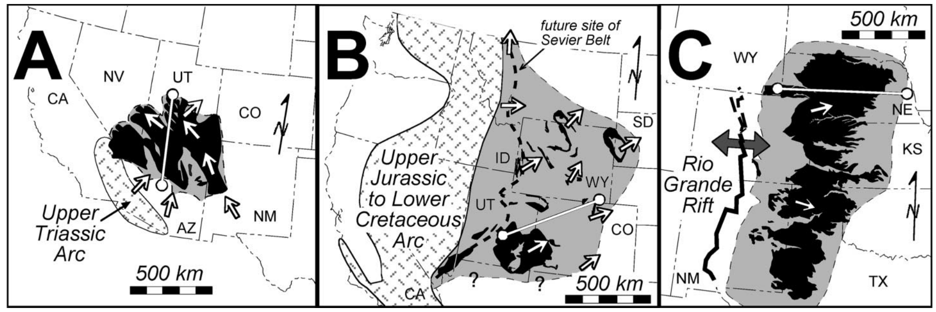

Figure 2. Maps of three gravel units: (A) Late Triassic time showing Shinarump Conglomerate Member of Chinle Formation (Stewart et al., 1972a; Blakey and Gubitosa, 1983; Dubiel and Brown, 1993); (B) Early Cretaceous time showing Cloverly Formation (in Wyoming) and its lithostratigraphic correlatives, including Buckhorn Member of Cedar Mountain Formation (in Utah), Lytle and Burro Mountain Formations (in Colorado), Lakota Formation (in South Dakota), and lower member of Ephraim Conglomerate (in Idaho and western Wyoming) (Heller and Paola, 1989); and (C) Miocene time showing Ogallala Formation of western Great Plains (Swinehart et al., 1985; Gustavson and Winkler, 1988). Maps include observed (black) and inferred (shaded) original distribution of units, generalized paleoflow trends of units (white arrows); line of reconstructed topography (Fig. 7) is shown as white line with round end points, and associated tectonic features are discussed in text (Baldridge et al., 1991; Armstrong and Ward, 1993; Chapin and Cather, 1994; Erslev, 2001). In Part B, heavy dashed line shows subsequent site of thrusting in Sevier belt. Generalized extension orientation of Rio Grande Rift is shown with double arrows and its interpreted northward extent (Tweto, 1979; Mears, 1993; Erslev, 2001) shown with dashed lines.

cal relief of up to 80 m along the basal unconformity suggests that the conglomerate filled paleovalleys, especially in its more upstream parts (Stewart et al., 1972a; Blakey and Gubitosa, 1983). Valley development diminishes downstream (northward), indicating that pre-Shinarump valley cutting was not likely due to sea-level change. Toward the north the unit is inferred to record deposition in a broad alluvial plain, and the preservation of mottled facies (Stewart et al., 1972a) is interpreted to represent soil formation in low-lying interfluves (Hasiotis et al., 2000).

Deposition of the Shinarump Conglomerate was penecontemporaneous with the onset of intrusive igneous activity in southeastern California and southern Arizona (Armstrong and Ward, 1993). The resulting Upper Triassic plutons are interpreted to represent the beginnings of Andean-type arc magmatism in the southwestern USA (Armstrong and Ward, 1993; Dickinson, 2000). The coincidence in timing of Shinarump deposition, at the base of the ash-rich Chinle Formation, and the onset of arc magmatism suggests to us that these events are linked.

Lower Cretaceous Conglomerates of the Rocky Mountains

Discontinuously exposed throughout the U.S. Rocky Mountains are conglomeratic fluvial rocks of Early Cretaceous age (Figs. 2B and 3). These units are called a variety of names, including the Buckhorn Member of the Cedar Mountain Formation in central Utah, the Burro Canyon and Lytle formations in Colorado, the Cloverly Formation in Wyo-

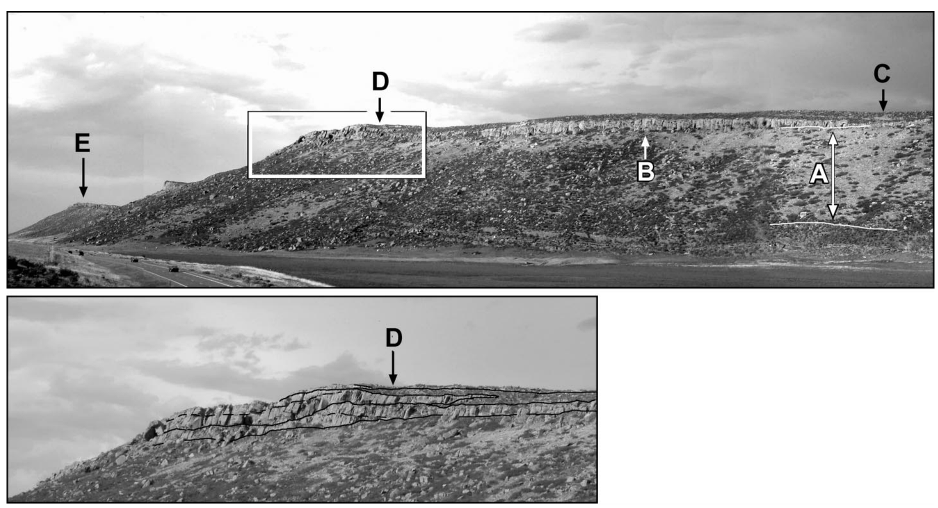

Figure 3. Example of widespread conglomeratic unit, in this case Lytle Formation (part of Lower Cretaceous gravels) north of Fort Collins, Colorado. This unit shows most typical characteristics of gravel units; it is conglomerate and sandstone, which abruptly overlies much finer-grained sedimentary deposits, in this case fluvial rocks of Late Jurassic Morrison Formation (A). Unit has sharp, disconformable contact with underlying unit (B). In most places, unit is a single, sheet-like deposit that pinches and swells (C), but there is evidence of amalgamation of several channel-belt bodies (D), enlarged below with superimposed line drawing. Resistant deposits typically form a cliff-capping unit that can be traced over many kilometers (E). Cars at base of cliff show scale. At D (inset), unit consists of three amalgamated sand/gravel bodies, of which upper two interfinger with finer-grained overbank deposits.

ming, and the Lakota Formation in the Black Hills. All of these units appear to sit disconformably upon the Morrison Formation (Fig. 3B), a widespread, mud-rich alluvial unit of Late Jurassic age (Stokes, 1944; Kowallis and Heaton, 1987). The duration and significance of the bounding unconformity has been contested (Furer, 1970; Winslow and Heller, 1987; May et al., 1995) and may represent only a lithofacies change in otherwise continuous deposition. The Lower Cretaceous conglomerates have been correlated to each other on the basis of composition, lithostratigraphy, grain-size trends, and general consistency of limited age dates (Heller and Paola, 1989). Some time-transgressive deposition of the conglomerate has been suggested based upon stratigraphic interpretations (Currie, 1997). Gravel-clast composition is primarily chert pebbles. Lithofacies interpretations suggest a low-sinuosity, braided origin (Heller and Paola, 1989; Zaleha et al., 2001).

The gravel body consists of amalgamated channels that locally include some finer-grained alluvial deposits (e.g., Fig. 3D; May et al., 1995). Overall, paleocurrent trends indicate flow toward the northeast (Heller and Paola, 1989), although some notable complexities are found locally, particularly in the Black Hills region (Zaleha et al., 2001). Age control is limited, but available biostratigraphic control suggests deposition during middle Neocomian to Aptian time (Tschudy et al., 1984; Heller and Paola, 1989).

Most accounts interpret the gravels as having been deposited prior to, and perhaps immediately prior to, the earliest deformation along the Sevier fold-thrust belt (Heller and Paola, 1989; Yingling and Heller, 1992; Currie, 1998). One interpretation suggests deposition in an overfilled earlier foreland basin that formed in front of thrust faults in central Nevada (DeCelles and Currie, 1996; Currie, 1998); however, the ages of thrust faults in that region are now thought to be Early to middle Cretaceous (MacCready et al., 1997), too young to generate an early foreland basin. In addition, if the gravels formed in a back-bulge basin setting, as suggested by Currie (1998) and DeCelles and Currie (1996), then a thick sequence of nonmarine foreland basin fill must have existed between the farthest west exposures of gravel and the inferred thrust belt that no longer exists. Considering that such a basin must have been very thick, in excess of 2 km thick, to develop a back bulge slope of 10^{-3} that is required by paleohydraulic analysis (Heller and Paola, 1989; Fig. 4), its complete absence is even more difficult to understand. We suggest that no such

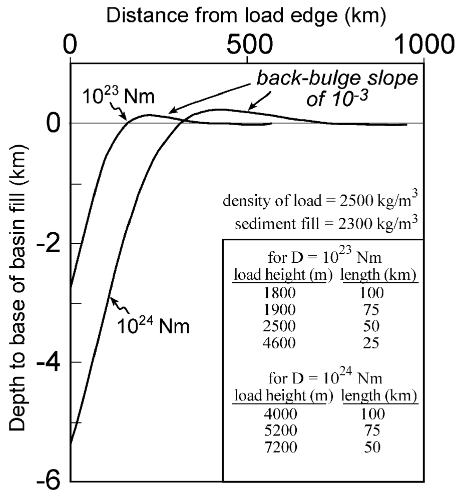

Figure 4. Flexural analysis of Lower Cretaceous gravels. Shown are filled sedimentary basin geometries required to generate back-bulge transport slopes of 10^{-3}, assuming an unbroken plate with flexural rigidities (D) of 10^{23} to 10^{24} Nm and for various load geometries (height and length). Width and depth of filled foreland basins associated with these back bulge slopes range from 200 km and 2.7 km deep to 400 km and 5.2 km deep, depending upon value of D.

"phantom foredeep" existed (Royse, 1993), but rather the region was simply dipping gently down toward the east.

Tectonic activity in the western U.S. during early Early Cretaceous time (Neocomian) was relatively quiescent (Dickinson, 2001). There is a notable paucity of sedimentary accumulation from that time until the initiation of the Sevier orogenic belt in Aptian/Albian time (Heller et al., 1986). The U.S. Cordilleran continental arc attained its maximum north-south extent in Jurassic time (Armstrong and Ward, 1993). During Late Jurassic time, the arc migrated eastward into westernmost Utah and had begun to retract westward during Early Cretaceous time. Apparently little else of regional tectonic significance took place, at least within or east of the volcanic arc, during the Early Cretaceous. Thus, the gravel was deposited during continued growth of the continental arc.

Ogallala Group

The Ogallala Group is part of a late Tertiary basin-fill sequence deposited within central Rocky Mountain basins and on the western Great Plains (Fig. 2C). In western Nebraska, geologists recognize several formations within the Ogallala Group, but lithologic and geometric complexities make broader correlation very difficult (Swinehart et al., 1985). The group is composed of predominantly fine-grained amalgamated fluvial and eolian deposits with dispersed interbedded conglomerates. Deposition began in the middle Miocene and continued until the earliest Pliocene (19–5 Ma) based on K/Ar and fission-track dating (Izett, 1975; Naeser et al., 1980) and biostratigraphic controls (Reeves, 1984; Swinehart et al., 1985). The Ogallala Group sits disconformably upon the Arikaree and White River groups in its northern exposure and unconformably on Mesozoic and Paleozoic rocks to the south. The unit marks the regional transition from a period of nearly continuous aggradation to one of net degradation (Diffendal, 1982; Swinehart et al., 1985).

Fluvial conglomeratic deposits of the Ogallala Group are remarkable in that they have been transported over 250 km from their source areas. These deposits form irregular cut-and-fill sequences that are most continuous in the Ash Hollow Formation along the Cheyenne Tablelands in eastern Wyoming and west-central Nebraska. They were deposited by east-northeast-flowing, low-sinuosity braided streams draining the Rocky Mountain front (Goodwin and Diffendal, 1987). To the south, conglomerates are less evident at the surface, but they are found in the subsurface (Gustavson and Winkler, 1988). Gravel compositions include granite and anorthosite clasts from the Front and Laramie ranges and volcanic clasts from flows in north-central Colorado (Blackstone, 1975). Relief on the base of the unit is up to 100 m, reflecting paleovalleys in the upstream end and local post-depositional subsidence due to salt dissolution in underlying Mesozoic units (Gustavson and Winkler, 1988; Oldham, 1996). The cut-and-fill alternations may be the result of climate fluctuations, autocyclic events, or tectonic uplift in the Rocky Mountains during the late Cenozoic.

Most interpretations of the Ogallala Group suggest that it records uplift associated with the propagating Rio Grande Rift (Trimble, 1980; Eaton, 1986). Deposition of the Ogallala Group occurred at the same time as deformation along the rift that was initiated in New Mexico between 28 and 27 Ma, reached central Colorado by 26–25 Ma (Chapin and Cather, 1994), and is manifested as 10–8 Ma volcanic rocks, post-Miocene normal faulting, and high heat-flow on the Colorado–Wyoming border (Keller and Baldridge, 1999; Mears, 1998). The link between the two is further supported by paleohydraulic analysis of Ogallala fluvial channels on the Cheyenne Tablelands.

These channels have been post-depositionally tilted up by an order of magnitude, most likely by tectonic uplift in the Rocky Mountains (McMillan et al., 2002). Thus, the Ogallala Group seems to be an indicator of long-wavelength tectonic activity both during and after its deposition.

Shared Characteristics

These three fluvial conglomerate units share several key features in common. While exposures today are discontinuous, paleogeographic reconstructions augmented by well data suggest that these units were far more continuous in the past and represent throughgoing alluvial systems. The Shinarump Conglomerate and Ogallala Group are dominantly sandstone, but gravels are found in locally abundant patches over long down-flow distances. The Lower Cretaceous units are more strongly conglomeratic. We think of these as discontinuous gravel sheets; however, we note that: (1) the dispersed presence of low-relief interfluves and occasional intercalated overbank deposits demonstrate that they formed as an amalgam of individual channel belts (Fig. 3D), and (2) there are lateral variations in grain size, particularly for the Shinarump Conglomerate and Ogallala Group, that demonstrate that a variety of individual river channel systems were active intermittently in space and time across the region. Most remarkably, the units are thin, typically a few meters thick, over nearly their entire lateral extent (Fig. 3). However, in their upstream areas, they cut and then fill paleovalleys into underlying deposits. The Shinarump Conglomerate contains many such paleovalleys toward the south that cut down into the Moenkopi Formation (Triassic) and the DeChelley Sandstone (Permian). These range up to a few tens of meters deep and up to 5 km wide, but northward the unit is more continuous, typically 10 m thick (up to 20 m in places) (Stewart et al., 1972a). The Lower Cretaceous gravels in Utah reach up to 35 m thick along trunk stream paleovalleys cut into the underlying Morrison Formation but are typically more consistent (~10 m) in thickness (Heller and Paola, 1989; Yingling and Heller, 1992; Currie, 1998). The entire Ogallala Group, deposited between 19 and 5 Ma (Swinehart et al., 1985), ranges from 30 to 240 m thick, however much of this fill is sandstone and the greatest thickness is confined to paleovalleys (Swinehart et al., 1985; Gustavson and Winkler, 1988; Gustavson, 1996).

The deposits contain clasts of resistant lithologies, such as quartz, quartzite, chert, igneous, and metamorphic pebbles, ranging up to cobble size, that were deposited by low-sinuosity (braided?) streams (Stokes, 1950; Blakey and Gubitosa, 1983; Swinehart et al., 1985; Gustavson and Winkler, 1988; Heller and Paola, 1989; Currie, 1998). Gravel bodies are widespread, laterally persistent, albeit not always continuous, and contain coarse-tail upward fining and/or nested channel cut-and-fill structures. Occasionally, the gravel units contain bar accretion surfaces, trough cross-bedding in the sandier fractions, and clast imbrication.

The units rest on underlying deposits with sharp erosional contact (Fig. 3B), which suggests that they are not parts of a continuous coarsening-upward progression but either flowed out upon a preexisting erosion surface of low relief or that they first beveled a low-angle scour surface (Stokes, 1950; Swinehart et al., 1985). The underlying units are dominantly fine-grained fluvial to nearshore marine deposits. Transport directions for units beneath the Lower Cretaceous gravels (the Morrison Formation) and Ogallala Group (Arikaree Group) were subparallel to the sheet gravels (Peterson, 1984, 1987; Winslow and Heller, 1987; Yingling and Heller, 1992). These underlying units are much finer grained than the gravel sheets, suggesting that they were transported by rivers of comparatively low slope. In the case of the Shinarump Conglomerate, the underlying Moenkopi Formation was not only a finer-grained, hence low-gradient, coastal plain to deltaic system, but paleoflow was at acute angles to that of the overlying gravel deposit (Stewart et al., 1972b). Thus, in all three cases, the onset of gravel deposition coincided with, or soon followed, a change to steeper transport gradients.

METHODOLOGY

Our approach combines two types of information. The first is a determination of the transport slope needed to move gravels out the observed distances from the source areas. River transport can be achieved by either building a sufficient slope or increasing flow depth to generate sufficient shear stresses to carry material. If we can constrain the flow depths from field observations, we can calculate the paleoslopes, which then record regional paleotopography.

The second type of information is unit thickness. The preserved thickness of gravel deposits and related units constrains how much aggradation can account for building the necessary topography to reach the needed transport slopes. The way these slopes are built comes from a combination of aggradation in the proximal part of basins, degradation in the distal parts of basins (not important in the case of net deposition), and tilting of topography by isostatic processes.

Paleoslope Estimates

The key to determining tilt histories is estimating paleoslopes from these widespread alluvial gravels. Various quantitative paleohydraulic analyses exist (e.g., Cotter, 1971; Ethridge and Schumm, 1978; Gardner, 1983); however, the most appropriate approach to determining paleoslopes for these types of coarse-grained deposits is that described by Paola and Mohrig (1996). This method utilizes the observed relationship between calculated basal shear stress and characteristic grain size of the mobile bed in gravelly braided streams. These streams tend to adjust to an equilibrium condition where basal shear stress is slightly above that needed to move a representative grain size. Assuming grain density of 2700 kg m^{-3}, slope (S) is related to grain size (D) and flow depth (H) by:

$$S = \psi D_x H^{-1}$$

(see Paola and Mohrig (1996) for derivation), where $\psi = 9.4 \times 10^{-2}$ when D_{50} (median diameter) is used (Paola and Mohrig, 1996), and $\psi = 2.38 \times 10^{-2}$ when D_{90} (90th percentile diameter) is used (Heller and Paola, 1989).

The slope calculation requires a measurement of characteristic flow depth (H) along reaches in which the measured gravels were deposited. We have chosen thickness of the fining-upward sequence in which gravels were measured as a proxy for flow depth. We realize that a preserved fining-upward sequence may vary slightly from true flow depth due to channel perching during filling, which would lengthen the fining-upward sequence, or subsequent scouring which might remove the upper part of the sequence. Although these effects tend to counter each other, we made an effort to reduce the second problem by only choosing sites where sequences include the fine-grained sandy to silty capping deposits.

Application of this method to modern gravelly rivers indicates uncertainties (2σ) of a factor of two or less (Paola and Mohrig, 1996). The four key assumptions in using this method are that flows were quasi-steady state, channel banks were noncohesive, bedforms were absent on the channel bed, and transport was dominantly as bedload. Thus, application is best applied to deposits that do not show signs of highly variable, short-lived, rapid deposition, where there is no evidence of plant

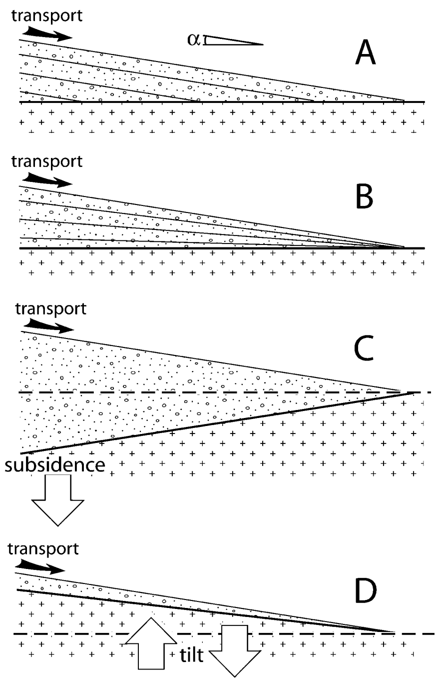

Figure 5. Conceptual model showing methodology used in this study. If paleotransport slope (α) and aggradational thickness of deposits (stippled pattern) are known, then tilt of underlying basement (patterned) can be deduced. (A) Required slope is built up atop horizontal basement. Slope is maintained during aggradation, leading to basinward progradation of system. (B) As in A, except transport slope steepens over time. This may result in a coarsening-upward sequence. (C) If thickness of unit increases upstream beyond that needed to maintain transport slope, then back-tilted subsidence must have occurred. This would be true in a typical foreland basin setting (Fig. 6). (D) If unit does not thicken upstream sufficiently to generate needed transport slope, then basement tilting must have occurred to reach transport slope α.

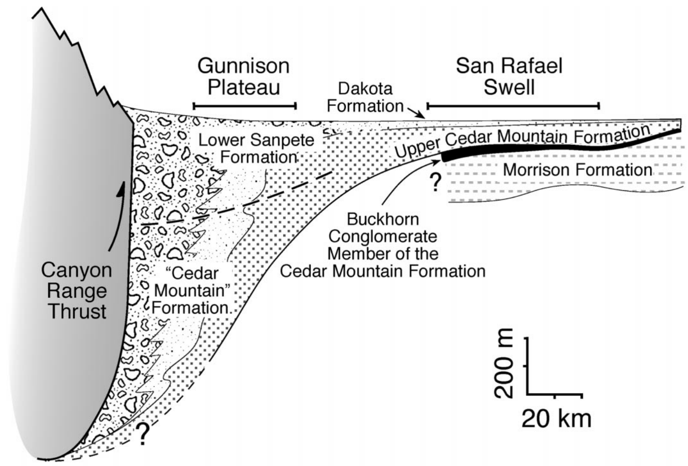

Figure 6. Interpreted cross section showing Early to middle Cretaceous units of central Utah that formed during initial growth of Sevier orogenic belt (modified from Yingling and Heller, 1992). Note strong asymmetry of basin fill and limited transport distance of conglomeratic deposits of "Cedar Mountain" and San Pete formations. Pre-foreland deposits include widespread gravel unit (in black)—Buckhorn Conglomerate Member of Cedar Mountain Formation of Early Cretaceous age.

roots or fine-grained cohesive sediment, and where channel-filling gravels are either massive or at most contain crude, parallel bedding. Field criterion for determining the suitability of deposits for analysis are more fully discussed in Paola and Mohrig (1996).

Table DR1 shows our calculations for paleoslopes at various sites for each unit.[2] Shinarump Conglomerate was collected at two sites in southern Nevada, along the east side of the Spring Mountains and at Firehole Valley (locations N-1 and N-3 of Stewart et al. [1972a]). Slopes for the Lower Cretaceous gravels were determined by Heller and Paola (1989) using D_{90} and come from eight sites between central Utah and southeastern Wyoming. Data from the Ogallala Group come from 10 sites across the Cheyenne Tablelands, which run east from southern Wyoming into western Nebraska (McMillan et al., 2002).

Our results indicate that paleoslopes for these transport systems span the range of 10^{-3} to 10^{-4}, not unlike modern river gradients for gravelly braided streams (Church and Rood, 1983; Paola and Mohrig, 1996). Estimated uncertainties on these values are about a factor of 2.

Determining Tilt

For the streams to reach and maintain the calculated paleoslopes, sufficient topography must have developed prior to or during deposition. One way topography can be developed is by higher aggradation rates in the upstream reaches, such as occurs during alluvial fan growth (Fig. 5A). In this case, early alluvial deposition takes place preferentially upstream and builds out over time; otherwise, deposition occurs basinwide but faster upstream so that a sufficiently steep transport slope forms. In the former case, sedimentation would downlap basinward over time. In the latter case, slopes would gradually increase upsection as relief increased. In either case, deposition must be greater upstream for relief to develop to the point where slopes could carry the observed gravels far into the basins (Fig. 5, A and B).

If more sedimentation takes place upstream than is needed to achieve the necessary slopes to carry the gravel forth, there must be subsidence in the proximal part of the basin to trap the extra sediment. This situation can be observed by hanging the available sedimentary thickness from the calculated paleoslope and seeing if the basement surface on which these deposits are laid had a back-tilting geometry (Fig. 5C). This case exists in most synorogenic gravel deposits (e.g. Fig. 6), where flexural or structural back tilting accompanies source-area uplift.

The converse is true when the aggradational thickness upstream is not sufficient to generate the needed slope to deliver gravels far out into the basin (Fig. 5D). In this case, the alternative way of generating enough topography to reach the needed transport slope is by tilting the basement down toward the distal part of the basin. The amount of tilting required is the difference between the maximum preserved thickness and the required slope. This tilting can be preexisting or can occur early during deposition of the gravel deposits.

A more complex alternative case is where transport slopes are formed by aggradation in the proximal basin followed by tilting and erosion in the same area after deposition of the gravels. Such is the case for the post-orogenic rebound model of Heller et al. (1988) and Flemings and Jordan (1989). However, in this

[2]GSA Data Repository item 2003121, data used in paleoslope reconstructions, is available on the Web at http://www.geosociety.org/pubs/ft2003.htm. Requests may also be sent to editing@geosociety.org.

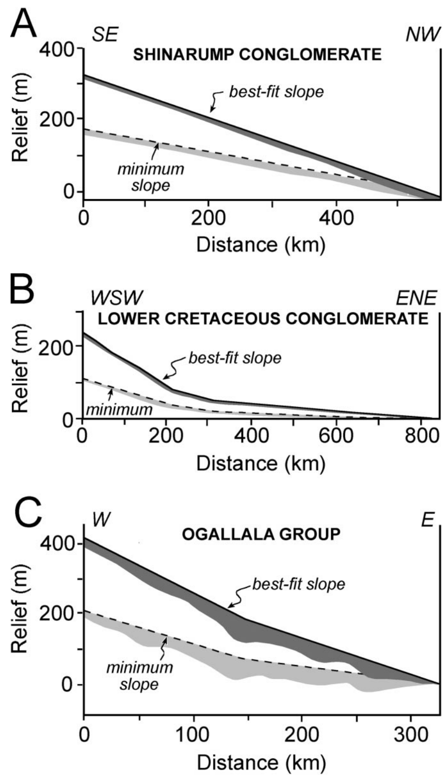

Figure 7. Results for three units: (A) Shinarump Conglomerate, with average slope of 6 × 10^{-4}; (B) Lower Cretaceous conglomerate, with average slope of 1.6 × 10^{-3} to west flattening to 2 × 10^{-4} eastward; (C) Ogallala Group (Ash Hollow Member) on Cheyenne Tablelands with average slope of 1.5 × 10^{-3} to west and 1 x 10^{-3} eastward. Generalized thicknesses of units are hung from this surface. Thickness is smoothed from measured sections and isopach maps reported in Stewart et al. (1972a) for Shinarump Conglomerate, Heller and Paola (1989) for Lower Cretaceous gravels, and Nebraska Geological Survey (1969) for Ogallala Group. Two transport slopes are shown. Best-fit slope (solid line) and minimum transport slope (dashed line; a factor of 2 less than best-fit slope). Lines of cross sections onto which data are projected are shown in Figure 2.

case, a significant unconformity would form atop the gravel units in the proximal part of the basin (Heller et al., 1988; Flemings and Jordan, 1989), which does not seem to be the case in our examples.

RESULTS

Figure 7 shows the results of our analysis. For each unit studied, the calculated topography of the river systems is shown as relief above the most downstream known extent of the gravel units. Topography is determined by projecting slopes upstream from these downstream points. Since we take topographic gradient to represent depositional slopes existing at the end of aggradation, we hang the preserved stratigraphic thickness of the gravelly units from this datum. In this figure, we show the generalized maximum stratigraphic thickness using results from published measured sections (Stewart et al., 1972a; Heller and Paola, 1989; McMillan et al., 2002). Although the conglomerate units are locally eroded away at present, we show the deposits as continuous, which they likely were at the end of deposition. Generalizing thickness in this way treats the stratigraphy as a continuous sheet along the cross sections, while preservation is patchy in truth. In addition, since we are drawing an envelope larger than the reported thicknesses, we err on the side of overestimating stratigraphic thickness for these units. This is a conservative approach (i.e., provides a minimum estimate of tilt), since the more aggradation can account for building transport slopes, the less tectonic tilting is needed.

The results for all three units are the same: the preserved stratigraphic thickness of the gravelly units is nowhere near enough to develop a sufficient transport slope by aggradation alone (Fig. 7). This is true even if minimum paleoslopes are used (Fig. 7, dashed lines). The amount of topography required beyond that of the aggradational buildup by the streams themselves is several hundred meters vertically over distances of several hundred kilometers (i.e. a slope of 10^{-3}). To develop this relief without evidence of significant differential aggradation in the upstream side or degradation in the downstream parts of the system requires tilting of the substratum.

DISCUSSION

Our results suggest that topography must have formed prior to, or during, deposition of the gravel units for the slopes to have been built steeply enough to deliver gravel far into the basins. Whatever mechanism generates such slopes, it is apparently not a typical occurrence as evidenced by the rarity of such widespread gravelly deposits.

The misfit between required slopes and available sediment thickness is so large that second-order effects—such as differential compaction, modest erosion of the proximal basin deposits or how the sections are hung—are not contenders. If significant post-depositional erosion of the proximal basin occurred, then to explain the results, (1) there must have been several hundred meters of erosion in the proximal basin to remove enough gravel to negate the requirement of tilting, and there is no evidence of significant, or any, erosion for these units (Witkind, 1956; Yingling and Heller, 1992); and (2) the amount of erosion would have to be carefully adjusted across the basin so the resultant deposit was of uniform thickness to yield the present sheet-like thickness of the units.

While the thickness of the gravel deposits is insufficient to generate enough relief to allow such extensive progradation, it could be that steep slopes were generated by differential aggradation and/or erosion of the underlying sedimentary units. To generate topography in this way, aggradation would have to be concentrated in the proximal part of the basin and erosion in the distal part. This is opposite to the observed patterns of erosion into underlying deposits. Erosional paleovalleys are primarily found beneath the upstream limits of the gravel units and are minor or absent farther downstream.

Underlying units are fluvial to fluvio-deltaic, presumably deposited at lower slopes than the overlying gravel. This need not be true for fine-grained fluvial deposits, as transport slope can be set by sediment flux and not just grain size. In the absence of an estimate of gradients for underlying deposits, we suspect they were much lower than the overlying gravels, based upon: (1) their fine-grained nature, which does not require steep gradient for transport, (2) the absence of gravel lenses forming an overall coarsening-upward sequence, which would show that gradients were building over time to a point where gravel could be continuously transported by the evolving stream system, and (3) the presence of disconformity and paraconformity below the gravel units, indicating that gravel deposition commenced after a discrete, indeterminate period of erosion.

Thus, we interpret these results to indicate that sufficient slopes existed to transport the widespread gravels and that the most plausible explanation is tilting of the Earth's surface over long (hundreds of kilometers) wavelengths.

Role of Climate

Given the thin, abrupt nature of these deposits, it might seem like rapid climate change could play a role in their origin. The primary influence of climate on the origin of alluvial stratigraphy is through four mechanisms: (1) an increase in precipitation, and therefore water flux in river systems; (2) changing the grain-size distribution derived from the source area by altering the mechanisms of weathering; (3) changing vegetation patterns, which can affect bank strength and hydrograph shape; and (4) by forcing increased regional erosion that leads to isostatic rebound and tilting.

Changing water flux in rivers affects river capacity and competence, which can generate at least transient gravel transport events (Paola et al., 1992). However, increased water flux does not play a role in the origin of these widespread gravels. Flow depth is a key observation made in reconstructing paleoslope; therefore, any increase in water depth has already been accounted for in the paleoslope determination. In addition, in truly braided river systems, changes in discharge are compensated for by changing the number of active channels and, therefore, the width of the active braid plain. Such lateral variations are unimportant in generating the shear stress necessary to move gravel at any single point. If the sheet-like distribution of the units required multiple channel threads active at the same time, then high water discharge would be required. However, it is not clear whether these units formed by multiple channels active at the same time or by the amalgamation of deposits formed by fewer channels active over a sustained period.

The second possible influence of climate is on the grain-size distribution delivered to rivers from hill slopes. A change in weathering process could change the relative abundance of gravel derived from source areas. A significant increase in the input material grain size would change the sorting of resultant deposits. While this influences the availability of gravel, it does not mitigate the need for tilt.

The third possible influence of climate is on channel pattern. If channel braiding is primarily a result of reduction of bank cohesiveness (Parker, 1978), then climate can come into play by both decreasing the abundance of clay-sized material in the water load and by altering the type and abundance of vegetation along the channel banks. The cohesiveness of clay exerts primary influence on bank stability and, therefore, channel form (Schumm and Thorne, 1989). Climate can also influence vegetation types along the flood plain, which can influence bank stability through rooting behavior. None of this negates the necessity to generate sufficient slopes to transport the materials. The only way for this scenario to work is if steep slopes formed earlier than the gravel deposits and the transition into gravels recorded a grain-size change from unroofed sources, not the onset of steeper gradients. However, evaluating this possibility would require reconstructing the history of slope changes in underlying, finer-grained rocks, which was not done in this study and would require a different methodology.

Fourth, while climate-induced erosion and resultant isostatic rebound provides a possible link between climate and tilting (e.g., Molnar and England, 1990), in the case of the widespread gravels, the area of deposition was over wavelengths far larger than flexural response distances of earth lithosphere (Watts, 1992; McKenzie and Fairhead, 1997; Stewart and Watts, 1997). Thus, for this mechanism to play a role would require broadly distributed, but asymmetric, erosion along most of the length of the depositional basin. Tilting at distances far from the eroded source areas requires an unusual distribution of erosion across the basin such that some areas are eroded deeply enough to generate significant isostatic rebound, while immediately adjacent areas continue to transport and deposit sands and gravels that record this tilt. Such a scenario is not only unlikely and oxymoronic for aggradational settings, but this hypothesis has been tested and negated for the Ogallala Group (McMillan et al., 2002).

Finally, if these deposits were related to climate, it would be curious for such events to be found only at these times and in these locations and not more broadly in North America or the world. It seems too geographically limited.

Role of Source Area Lithology

If source areas contained a change in lithology either in depth or over an area, then a change in abundance of gravel production could reflect, respectively, simple unroofing or integration of streams into larger catchments. Either way, lithologies capable of generating more resistant gravel-sized clasts could be tapped over time. This is similar to the second possible climatic influence discussed above, and, as with that case, it has no bearing on the need to generate a tilt to allow these gravels to reach points farther out across the depositional basin. However, it does provide an intriguing possibility that tilting began earlier than we assume and that the occurrence of gravel reflects unroofing sometime after tilting had started. We suspect that such may be the case for the Lower Cretaceous gravels in the Rocky Mountains. The underlying unit, the Morrison Formation of Late Jurassic age, is fine-grained overall, but it does contain gravel along its most upstream, western, exposures (Peterson, 1980). In particular, the Salt Wash Member of the Morrison Formation is relatively coarse grained. Based on detrital fossils preserved within clasts of the Lower Cretaceous gravel (Winslow and Heller, 1987; Heller and Paola, 1989), the source area includes upper Paleozoic rocks of Nevada. These areas have a suitable stratigraphy for unroofing to play a major role in the evolution of the Late Jurassic through Early Cretaceous stratigraphy in the Rocky Mountains. Triassic rocks of Nevada contain many marine mud-rich units, including the Auld Lang Syne Group, often referred to informally as the ''mud-pile.'' These deposits sit atop Pennsylvanian-Permian age shelf carbonates that include secondary chert nodules. Erosion of this source area could lead to the dominantly mud-rich fluvial rocks of the Morrison Formation. As deeper levels were unroofed in the source area, the Paleozoic chert-bearing carbonates could be tapped, providing clasts found in the Lower Cretaceous gravels. If this is the case, it may well be that tilting began in Late Jurassic time in this area and helped generate slopes in the Morrison Formation. This possibility could be tested once a methodology for reconstructing slopes in finer-grained fluvial systems is established.

Tectonic Timing of Continental Tilting

In search of a causal mechanism for large-scale continental tilting, we note that each of the gravel sheets is associated with changes in mantle density structure associated either with subduction or late Cenozoic arrival of warmer-than-normal mantle beneath the southern Rocky Mountains. The Shinarump Conglomerate was deposited during Late Triassic time (Stewart et al., 1972a; Blakey and Gubitosa, 1983), coincident with a transition from a passive (and/or transform) margin to initiation of an Andean-type convergent margin in the southwestern USA (Walker, 1988; Dickinson, 2000). Continental arc plutonism began after deposition of the Moenkopi Formation and was commensurate with deposition of the Chinle Formation, of which the Shinarump is the basal member (Stewart et al., 1972a, b; Armstrong and Ward, 1993; Dickinson, 2000).

Lower Cretaceous gravels were deposited sometime during Neocomian-Aptian time, close to, but arguably immediately prior to, earliest thrusting in the Sevier belt (Fig. 2B) (Heller et al., 1986; Yingling and Heller, 1992; Currie, 1998). The onset of deformation in the Sevier belt post-dated deposition of the gravels, as determined from source area provenance of gravel composition (Heller and Paola, 1989; Yingling and Heller, 1992), lack of contemporaneous foreland basin subsidence

(Heller et al., 1986), and the known chronology of tectonic activity in adjacent parts of Nevada (Dickinson, 2001). Also, there is no geologically reasonable rendition of thrust belt loading that can result in basinward tilts of sufficient steepness or width to generate the observed distribution of Lower Cretaceous gravels (Fig. 4; c.f. Heller et al., 1986; DeCelles and Giles, 1996; Currie, 1998). More likely, basin partitioning by thrust-belt deformation led to local trapping of gravel deposits, such as seen in the early foreland basin fill of Utah (Fig. 6). Deposition of the Lower Cretaceous gravels instead took place following a period of continued eastward expansion of the Mesozoic continental arc in the western USA (Armstrong and Ward, 1993; Dickinson, 2001).

The Ogallala Group was deposited eastward across the western Great Plains following increased magmatic activity in Colorado and New Mexico (Chapin and Cather, 1994), uplift, and northward-propagation of the Rio Grand Rift. In New Mexico and southern Colorado, rifting began in Oligocene time (Chapin and Cather, 1994). Farther north, the surface manifestation of rifting is discontinuous, and available age data suggest a more recent initiation (Leat et al., 1991; Mears, 1993; Erslev, 2001). Young uplift of the western Great Plains, inferred to relate to propagation of the rift (Angevine and Flanagan, 1987), is coincident with deposition of the Ogallala Group (Miocene) and younger fluvial units in that area (McMillan et al., 2002). Symmetric deposition to the west of the rift was neither continuous nor widespread, presumably due to inherited topography from earlier orogeny that forced deposition into more isolated basins.

Tilt Origins

In each case, gravel sheet deposition mostly took place inboard of the locus of tectonic activity and was not directly affected by associated structural deformation; therefore, crustal-scale structure played no role in regional tilting. Tilting at wavelengths greater than a few hundred kilometers can be accomplished in a couple of ways. Isostatic topography results primarily from lithospheric density variations. A defining characteristic of isostatic topography is that it generally changes slowly in response to erosion and major orogenic/magmatic episodes. Dynamic topography, on the other hand, is the deflection of the Earth's surface that balances the stresses applied at the base of the elastic lithosphere by mantle convection (Burgess and Moresi, 1999). In contrast to isostatic topography, dynamic topography is transient because it results from the stresses created by the evolving pattern of mantle convection. Dynamic topography is calculated by solving the equations for mantle convection with a specified distribution of density and viscosity throughout the Earth's mantle. The effect of the 660 km phase transition on mass transport across this boundary can also significantly modulate the time evolution of dynamic topography, especially on short time scales (Pysklywec and Mitrovica, 2000).

While the magnitude of dynamic topography is model-dependent, it is now widely accepted that most lateral variations in mantle density structure can be reconstructed by integrating the last 180 m.y. of plate subduction (Richards and Engebretson, 1992). The remaining dynamic topography probably results from warm upwellings from the deep mantle (Lithgow-Bertelloni and Silver, 1998). In the case of the western USA, the subduction of 7000 km of oceanic plate over the last 180 Ma is well documented (Engebretson et al., 1992) and apparently has exerted a first-order control on the dynamic topography of the region as documented by the sedimentary record (Mitrovica et al., 1989; Coakley and Gurnis, 1995; Burgess et al., 1997; Pysklywec and Mitrovica, 2000). More controversial are the roles of tectonic uplift (due to increased mantle buoyancy) versus climate change in the southern Rocky Mountains as the cause of deposition of the post-Laramide Ogallala Group (McMillan et al., 2002).

Our observations suggest that 400–800 m of differential uplift occurred over areas 400–800 km wide on time scales of 1–10 m.y. These tilts are consistent with the prediction of dynamic topography from time-varying subduction models (Mitrovica et al., 1989; Lithgow-Bertelloni and Silver, 1998; Burgess and Moresi, 1999; Pysklywec and Mitrovica, 2000). While it is true that the dynamic topography is filtered by the lithosphere's elastic response at wavelengths shorter than 500–1000 km, we do not consider its effect important to our first-order conclusions.

Shinarump Conglomerate

While little is known about plate kinematics during Late Triassic time, we assume the Shinarump Conglomerate is related to the penecontemporaneous onset of subduction along the southwestern U.S. continental margin (Fig. 8A). Certainly substantial variations in isostatic and dynamic topography are predicted during the early evolution of any subduction zone as a result of initial slab emplacement, initiation of arc magmatism, and attendant changes in extensional-compressional regimes. Of the cases studied, the Shinarump Conglomerate is the least compelling for dynamic topography because the wavelength of deposition is smaller and the variability of paleocurrents is greater than in the other cases. However, the correlation of timing of deposition of the Shinarump following the onset of subduction is consistent with isostatic adjustments associated with arc creation playing a major role in regional tilting. More data on the timing of Late Triassic volcanism and tectonism in the southwestern USA and/or a better estimate of plate kinematics at the time are needed to constrain dynamical models.

Cretaceous Conglomerate

As the western U.S. subduction zone lengthened over the next few tens of millions of years, so too did the associated continental arc grow along the North American Cordillera (Armstrong and Ward, 1993). By Cretaceous time, the North American continent had drifted over the large mass of negatively buoyant subducted slab that had accumulated since Late Triassic time. The sinking of this mass creates lithospheric subsidence with relief of hundreds of meters over wavelengths of hundreds of kilometers (Mitrovica et al., 1989; Gurnis, 1991, 1992). These tilts, approaching 10^{-3}, are sufficient to promote the observed widespread distribution of Lower Cretaceous gravels (Fig. 8B). With continued westward migration of the continent over the moment of accumulated slab mass in the middle, mantle migrates beneath the central North American continent by Late Cretaceous time. This, coincident with a time of global sea level highstand (Haq et al., 1987), produces the subsidence needed to create the Western Interior Seaway (Fig. 8C) (Gurnis, 1992; Mitrovica et al., 1989). It is notable that topography inferred for the Western Interior Seaway (a few hundred meters of relief over a few hundred kilometers in length [Martinson et al., 1998]) is the same scale as is needed immediately preceding deposition of the Lower Cretaceous gravels somewhat farther west. While we suspect that other widespread fluvial units deposited between Triassic and Cretaceous time, specifically the Kayenta and Morrison formations, may be related to the same phenomena (e.g., Lawton, 1994), paleoslope reconstructions of these finer-grained units are not yet available.

Ogallala Group

Finally, we relate continental tilting associated with the Ogallala Group with the occurrence of the Neogene uplift of the southern and central Rockies (Fig. 8D) (Eaton, 1987; McMillan et al., 2002). This uplifted region has been correlated with low-velocity upper mantle currently imaged beneath this region that is interpreted to be mantle at or near the dry peridotite solidus (Goes and van der Lee, 2002). The North American upper mantle density model of Goes and van der Lee (2002) predicts ~1 km of isostatic elevation difference between the southern Rocky Mountains and Kansas. This effective tilt on the order of 10^{-3} is sufficient to enable transport of the gravel deposits of the Ogallala Group. We would suggest that a peak in volcanism in Colorado and New Mexico ca. 35 Ma (McMillan Nancy et al., 2000), coincident with the onset of extension along the Rio Grande Rift (Chapin and Cather, 1994), marks the onset of a lithospheric warming event associated with regional uplift. At longer wavelengths, the dynamical rebound of the western USA is predicted as the North American continent drifted well west of the Farallon slab, which now resides in the middle mantle off the east coast of North America (Grand et al., 1997). Thus, a combination of dynamic rebound and the regional lithospheric warming event are responsible for the origin of the Ogallala Group, as opposed to subsidence, which led to the formation of the other two gravel sheets. The preservation potential of the Ogalla Group is therefore low; continued uplift and tilting of the deposits as the rift continues to grow leads to erosion and reworking of the previously deposited materials (Heller et al., 1988).

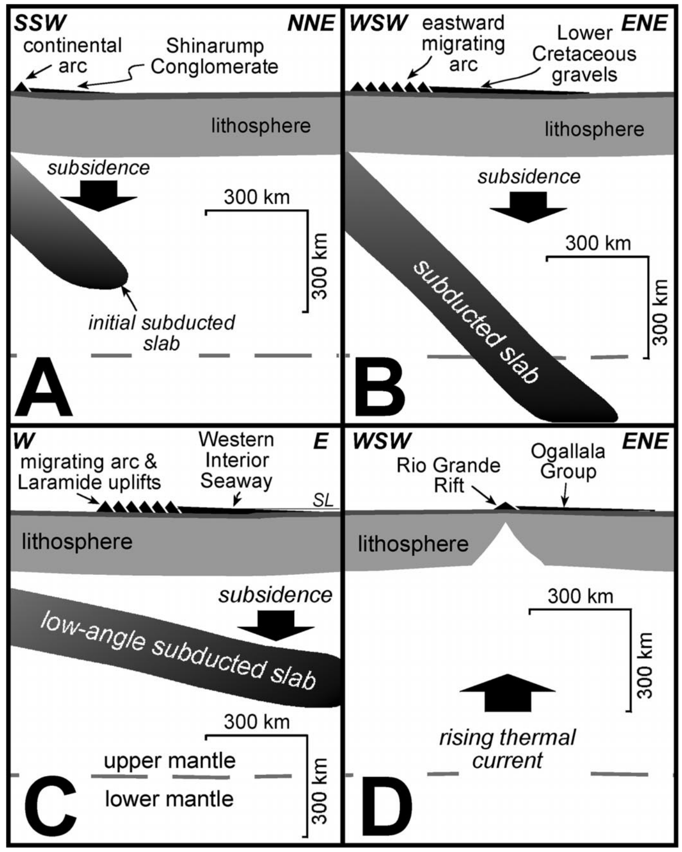

Figure 8. Interpreted relationship of mantle features to regional tilt and deposition of widespread units. (A) Subsidence associated with initial underthrusting of subducted slab is correlated with tilt and deposition of Shinarump Conglomerate in Late Triassic time. Penetration of slab may have caused subsidence either by inducing asthenospheric counterflow above subducted slab or by disrupting vertical heat loss from mantle. (B) Continued subduction, growth, and migration of continental arc is related to northeast-migrating tilt and depocenter recorded by Lower Cretaceous gravel units in Rocky Mountains. (C) Continued eastward migration of tilt and Western Interior Seaway depocenter of Late Cretaceous age is tied to gradual eastward migration of low-angle subducted slab. SL—sea level during Cretaceous eustatic highstand. (D) Tilting associated with progradation of Ogallala Group is interpreted to relate to regional uplift and active extension associated with evolving Rio Grande Rift in Miocene time.

CONCLUSIONS

By combining paleohydraulic requirements of gravel transport with preserved isopach geometries, we analyzed the tilt history of the western USA at various times in its evolution. The presence of widespread gravels, such as those described here, indicates large-scale deflections of continental crust in areas with little preexisting topography. Such regional continental tilts result from dynamic topography created by mantle convection. The rarity of these widespread units suggests that either periods of regional continental tilting are unusual or that preservation potential of these deposits is low. However, it is possible that similar thin, widespread conglomeratic units may have been recognized elsewhere but were lumped in with surrounding alluvial deposits for the purposes of geologic mapping. Identification of such units is important because they provide at least semiquantitative evidence of broad wavelength topography that can help constrain the time-integrated effect of slab subduction and mantle convection on surface processes. As such, paleotopography, as determined from calculated transport gradients of sedimentary deposits, provides a means of relating constructional landforms to mantle-driven geodynamics.

ACKNOWLEDGMENTS

Supported in part by National Science Foundation grant EAR-0106029. Helpful reviews were provided by R. Dorsey, E. Humphreys and D. Mohrig. We are grateful for discussions and/or field assistance provided by R.C. Blakey, M. Cheadle, W.R. Dickinson, R. Dunne, S. Liu, H. Luo, R.M. Lynds, C. Paola, D. Saffer, A.W. Snoke, A.K. Souders, and K. Trujillo.

REFERENCES CITED

Angevine, C.L., and Flanagan, K.M., 1987, Buoyant subsurface loading of the lithosphere in the Great Plains foreland basin: Nature, v. 327, p. 137–139.

Armstrong, R.L., and Ward, P.L., 1993, Late Triassic to Earliest Eocene magmatism in the North American Cordillera: Implications for the Western Interior Basin, *in* Caldwell, W.G.E., and Kauffman, E.G., eds., Evolution of the Western Interior Basin: Geological Association of Canada Special Paper 39, p. 49–72.

Baldridge, W.S., Perry, F.V., Vaniman, D.T., Nealey, L.D., Leavy, B.D., Laughlin, A.W., Kyle, P., Bartov, Y., Steinitz, G., and Gladney, E.S., 1991, Middle to late Cenozoic magmatism of the southeastern Colorado Plateau and central Rio Grande rift (New Mexico and Arizona, U.S.A.): A model for continental rifting: Tectonophysics, v. 197, p. 327–354.

Blackstone, D.L., Jr., 1975, Late Cretaceous and Cenozoic history of Laramie Basin region, southeast Wyoming: Geological Society of America Memoir 144, p. 249–279.

Blakey, R.C., and Gubitosa, R., 1983, Late Triassic paleogeography and depositional history of the Chinle Formation, southern Utah and northern Arizona, *in* Reynolds, M.W., and Dolly, E.D., eds., Mesozoic paleogeography of west-central United States: Denver, Colorado, Society of Economic Paleontologists and Mineralogists (SEPM) Rocky Mountain Paleogeography Symposium, v. 2, p. 57–76.

Burgess, P.M., and Moresi, L.N., 1999, Modelling rates and distribution of subsidence due to dynamic topography over subducting slabs: Is it possible to identify dynamic topography from ancient strata?: Basin Research, v. 11, p. 305–314.

Burgess, P.M., Gurnis, M., and Moresi, L., 1997, Formation of sequences in the cratonic interior of North America by interaction between mantle, eustatic, and stratigraphic processes: Geological Society of America Bulletin, v. 109, p. 1515–1535.

Chapin, C.E., and Cather, S.M., 1994, Tectonic setting of the axial basins of the northern and central Rio Grande Rift, *in* Keller, G.R., and Cather, S.M., eds., Basins of the Rio Grande Rift; structure, stratigraphy, and tectonic setting: Geological Society of America Special Paper 291, p. 5–25.

Church, M., and Rood, K., 1983, Catalogue of alluvial river channel regime data: Vancouver, University British Columbia, Department of Geography, 99 p.

Coakley, B., and Gurnis, M., 1995, Far-field tilting of Laurentia during the Ordovician and constraints on the evolution of a slab under an ancient continent: Journal of Geophysical Research, v. 100, p. 6313–6327.

Cotter, E., 1971, Paleoflow characteristics of a Late Cretaceous river in Utah from analysis of sedimentary structures in the Ferron Sandstone: Journal of Sedimentary Petrology, v. 41, p. 129–138.

Currie, B.S., 1997, Sequence stratigraphy of nonmarine Jurassic–Cretaceous rocks, central Cordilleran foreland-basin system: Geological Society of America Bulletin, v. 109, p. 1206–1222.

Currie, B.S., 1998, Upper Jurassic-Lower Cretaceous Morrison and Cedar Mountain Formations, NE Utah-NW Colorado: Relationships between nonmarine deposition and early Cordilleran foreland-basin development: Journal of Sedimentary Research, v. 68, p. 632–652.

DeCelles, P.G., and Currie, B.S., 1996, Long-term sediment accumulation in the Middle Jurassic-early Eocene Cordilleran retroarc foreland-basin system: Geology, v. 24, p. 591–594.

DeCelles, P.G., and Giles, K.A., 1996, Foreland basin systems: Basin Research, v. 8, p. 105–123.

Dickinson, W.R., 2000, Geodynamic interpretation of Paleozoic tectonic trends oriented oblique to the Mesozoic Klamath-Sierran continental margin in California, *in* Soreghan, M.J., and Gehrels, G.E., eds., Paleozoic and Triassic paleogeography and tectonics of western Nevada and northern California: Geological Society of America Special Paper 347, p. 209–245.

Dickinson, W.R., 2001, Tectonic setting of the Great Basin through geologic time: Implications for metallogeny, *in* Shaddrick, D.R., Zbinden, E.A., Mathewson, D.C., and Prenn, C., eds., Regional tectonics and structural control of ore: The major gold trends of northern Nevada: Geological Society of Nevada Special Publication 33, p. 27–53.

Diffendal, R.F., Jr., 1982, Regional implications of the geology of the Ogallala Group (upper Tertiary) of southwestern Morrill County, Nebraska and adjacent areas: Geological Society of America Bulletin, v. 93, p. 964–976.

Dubiel, R.F., and Brown, J.L., 1993, Sedimentologic analysis of cores from the Upper Triassic Chinle Formation and Lower Permian Cutler Formation, Lisbon Valley, Utah: U.S. Geological Survey Bulletin, v. 2000-E, p. E1–E40.

Eaton, G.P., 1986, A tectonic redefinition of the southern Rocky Mountains: Tectonophysics, v. 132, p. 163–193.

Eaton, G.P., 1987, Topography and origin of the southern Rocky Mountains and Alvarado Ridge, *in* Coward, M.P., Dewey, J.F., and Hancock, P.L., eds., Continental extensional tectonics: Geological Society [London] Special Paper 28, p. 355–369.

Engebretson, D.C., Kelley, K.P., Cashman, H.J., and Richards, M.A., 1992, 180 million years of subduction: Geological Society of America Today, v. 2, p. 93–100.

Erslev, E.A., 2001, Multistage, multidirectional Tertiary shortening and compression in north-central New Mexico: Geological Society of America Bulletin, v. 113, p. 63–74.

Ethridge, F.G., and Schumm, S.A., 1978, Reconstructing paleochannel morphologic and flow characteristics: Methodology, limitations and assessment, *in* Miall, A.D., ed., Fluvial Sedimentology: Calgary, Canadian Society of Petroleum Geologists Memoir 5, p. 703–721.

Flemings, P.B., and Jordan, T.E., 1989, A synthetic stratigraphic model of foreland basin development: Journal of Geophysical Research, v. 94, p. 3851–3866.

Furer, L.C., 1970, Petrology and stratigraphy of nonmarine Upper Jurassic-Lower Cretaceous rocks of western Wyoming and southeastern Idaho: American Association of Petrology Geology Bulletin, v. 54, p. 2282–2302.

Gardner, T.W., 1983, Paleohydrology and paleomorphology of a Carboniferous, meandering, fluvial sandstone: Journal of Sedimentary Petrology, v. 53, p. 991–1005.

Goes, S., and van der Lee, S., 2002, Thermal structure of the North American uppermost mantle inferred from seismic tomography: Journal of Geophysical Research, v. 107, p. 2000JB000049.

Goodwin, R.G., and Diffendal, R.F., Jr., 1987, Paleohydrology of some Ogallala (Neogene) streams in the southern panhandle of Nebraska, *in* Ethridge, F.G., et al., ed., Recent developments in fluvial sedimentology, Volume 39: Society of Economic Paleontologists and Mineralogists Special Publication: Tulsa, Society of Economic Paleontologists and Mineralogists (SEPM), p. 149–157.

Grand, S.P., van der Hilst, R.D., and Widiyantoro, S., 1997, Global seismic tomography: A snapshot of convection in the Earth: Geological Society of America Today, v. 7, p. 1–7.

Gurnis, M., 1991, Continental flooding and mantle-lithosphere dynamics, *in* Sabadini, R., Lambeck, K., and Boschi, E., eds., Glacial isostasy, sea-level and mantle rheology: Dordrect, Kluwer Academic Publishers, p. 445–491.

Gurnis, M., 1992, Rapid continental subsidence following the initiation and evolution of subduction: Science, v. 255, p. 1556–1558.

Gustavson, T.C., 1996, Fluvial and eolian depositional systems, paleosols, and paleoclimate of the Upper Cenozoic Ogallala and Blackwater Draw Formations, southern high plains, Texas and New Mexico: Austin, The University of Texas, 62 p.

Gustavson, T.C., and Winkler, D.A., 1988, Depositional facies of the Miocene-Pliocene Ogallala Formation, northwestern Texas and eastern New Mexico: Geology, v. 16, p. 203–206.

Haq, B.U., Hardenbol, J., and Vail, P.R., 1987, Chronology of fluctuating sea levels since the Triassic: Science, v. 235, p. 1156–1167.

Hasiotis, S.T., Demko, T.M., and Dubiel, R.F., 2000, Shinarump incised valley cut-and-fill deposits, paleosols, and ichnofossils in the lower part of the Upper Triassic Chinle Formation, Petrified Forest National Park, Arizona: It really does exist, and it has regional stratigraphic implications: Geological Society of America Abstracts with Programs, v. 33, p. 23.

Heller, P.L., and Paola, C., 1989, The paradox of Lower Cretaceous gravels and the initiation of thrusting in the Sevier orogenic belt, United States Western Interior: Geological Society of America Bulletin, v. 101, p. 864–875.

Heller, P.L., Angevine, C.L., Winslow, N.S., and Paola, C., 1988, Two-phase stratigraphic model of foreland-basin sequences: Geology, v. 16, p. 501–504.

Heller, P.L., Bowdler, S.S., Chambers, H.P., Coogan, J.C., Hagen, E.S., Shuster, M.W., Winslow, N.S., and Lawton, T.F., 1986, Time of initial thrusting in the Sevier orogenic belt, Idaho-Wyoming and Utah: Geology, v. 14, p. 388–391.

Izett, G.A., 1975, Late Cenozoic sedimentation and deformation in northern Colorado and adjoining areas, *in* Curtis, B.F., ed., Cenozoic history of the Southern Rocky Mountains: Geological Society of America Memoir 144: Boulder, Geological Society of America, p. 179–209.

Keller, G.R., and Baldridge, W.S., 1999, Lithospheric structure and evolution of the Rocky Mountains, Part II: Rocky Mountain Geology, v. 34, p. 121–130.

Kowallis, B.J., and Heaton, J.S., 1987, Fission-track dating of bentonites and bentonitic mudstones from the Morrison Formation in central Utah: Geology, v. 15, p. 1138–1142.

Lawton, T.F., 1994, Tectonic setting of Mesozoic sedimentary basins, Rocky Mountain Region, United States, *in* Caputo, M.V., Peterson, J.A., and Franczyk, K.J., eds., Mesozoic systems of the Rocky Mountain Region, Society for Sedimentary Geology (SEPM) Rocky Mountain Section, p. 1–25.

Leat, P.T., Thompson, R.N., Morrison, M.A., Hendry, G.L., and Dickin, A.P., 1991, Alkaline hybrid mafic magmas of the Yampa area, NW Colorado, and their relationship to the Yellowstone mantle plume and lithospheric mantle domains: Contributions to Mineralogy and Petrology, v. 107, p. 310–327.

Lithgow-Bertelloni, C., and Silver, P.G., 1998, Dynamic topography, plate driving forces and the African superswell: Nature, v. 395, p. 269–272.

Lucas, S.G., 1993, The Chinle Group: Revised stratigraphy and biochronology of Upper Triassic nonmarine strata in the western United States: Museum of Northern Arizona Bulletin, v. 59, p. 27–50.

MacCready, T., Snoke, A.W., Wright, J.E., and Howard, K.A., 1997, Mid-crustal flow during Tertiary extension in the Ruby Mountains core complex, Nevada: Geological Society of America Bulletin, v. 109, p. 1576–1594.

Martinson, V.S., Heller, P.L., and Frerichs, W.E., 1998, Distinguishing middle Late Cretaceous tectonic events from regional sea-level change using foraminiferal data from the U.S. Western Interior: Geological Society of America Bulletin, v. 110, p. 259–268.

May, M.T., Furer, L.C., Kvale, E.P., Suttner, L.J., Johnson, G.D., and Meyers, J.H., 1995, Chronostratigraphy and tectonic significance of Lower Cretaceous conglomerates in the foreland of central Wyoming, *in* Dorobek, S.L., and Ross, G.M., eds., Stratigraphic evolution of foreland basins: Tulsa, Oklahoma, Society for Sedimentary Geology (SEPM) Special Publication 52, p. 97–110.

McKenzie, D., and Fairhead, D., 1997, Estimates of the effective elastic thickness of the continental lithosphere from bouguer and free air gravity anomalies: Journal of Geophysical Research, v. 102, p. 27,523–27,552.

McMillan, M.E., Angevine, C.L., and Heller, P.L., 2002, Postdepositional tilt of the Miocene-Pliocene Ogallala Group in the western Great Plains: Evidence of late Cenozoic uplift of the Rocky Mountains: Geology, v. 30, p. 63–66.

McMillan, Nancy, J., Alan, P.D., and Haag, D., 2000, Evolution of magma source regions in the Rio Grande Rift, southern New Mexico: Geological Society of America Bulletin, v. 112, p. 1582–1593.

Mears, B., Jr., 1993, Geomorphic history of Wyoming and high-level erosion surfaces, *in* Snoke, A.W., Steidtmann, J.R., and Roberts, S.M., eds., Geology of Wyoming: Laramie, Geological Survey of Wyoming Memoir 5, p. 698–626.

Mears, B., Jr., 1998, Neogene normal faulting superposed on a Laramide uplift: Medicine Bow Mountains, Sierra Madre, and intervening Saratoga Valley, Wyoming and Colorado: Contributions to Geology, v. 32, p. 181–185.

Mitrovica, J.X., Beaumont, C., and Jarvis, G.T., 1989, Tilting of continental interiors by the dynamical effects of subduction: Tectonics, v. 8, p. 1079–1094.

Molnar, P., and England, P., 1990, Late Cenozoic uplift of mountain ranges and global climate change: Chicken or egg?: Nature, v. 346, p. 29–34.

Naeser, C.W., Izett, G.A., and Obradovich, J.D., 1980, Fission-track and K-Ar ages of natural glasses: U.S. Geological Survey Bulletin 1489, 31 p.

Nebraska Geological Survey, 1969, Geologic bedrock map of Nebraska: Lincoln, Nebraska Geological Survey, scale 1:2,900,000, 1 sheet.

Oldham, D.W., 1996, Permian salt in the northern Denver Basin; controls on occurrence and relationship to oil and gas production from Cretaceous reservoirs, *in* Longman, M.W., and Sonnenfeld, M.D., eds., Paleozoic systems of the Rocky Mountain region: Denver, Society for Sedimentary Geology (SEPM), Rocky Mountain Section, p. 335–354.

Paola, C., and Mohrig, D., 1996, Paleohydraulics revisited: Paleoslope estimation in coarse-grained braided rivers: Basin Research, v. 8, p. 243–254.

Paola, C., Heller, P.L., and Angevine, C.L., 1992, The large-scale dynamics of grain-size variation in alluvial basins, 1: Theory: Basin Research, v. 4, p. 73–90.

Parker, G., 1978, Self-formed rivers with equilibrium banks and mobile bed. Part 2: The gravel river: Journal of Fluid Mechanics, v. 89, p. 127–146.

Peterson, F., 1980, Sedimentology of the uranium-bearing Salt Wash Member and Tidwell unit of the Morrison Formation in the Henry and Kaiparowits Basins, Utah, *in* Picard, M.D., ed., Henry Mountains Symposium: Salt Lake City, Utah Geological Association Publication, v. 8, p. 305–322.

Peterson, F., 1984, Fluvial sedimentation on a quivering craton: Influence of slight crustal movements on fluvial processes, Upper Jurassic Morrison Formation, western Colorado Plateau: Sedimentary Geology, v. 38, p. 21–49.

Peterson, F., 1987, The search for source areas of Morrison (Upper Jurassic) clastics on the Colorado Plateau: Geological Society of America Abstracts with Programs, v. 19, p. 804.

Pysklywec, R.N., and Mitrovica, J.X., 2000, Mantle flow mechanisms of epeirogeny and their possible role in the evolution of the western Canada sedimentary basin: Canadian Journal of Earth Sciences, v. 37, p. 1535–1548.

Reeves, C.C., Jr., 1984, The Ogallala depositional mystery, *in* Whetstone, G.A., ed., Proceedings, Ogallala Aquifer symposium II: Lubbock, Texas Tech University Water Resources Research Center, p. 129–156.

Richards, M.A., and Engebretson, D.C., 1992, Large-scale mantle convection and the history of subduction: Nature, v. 355, p. 437–440.

Riggs, N.R., Lehman, T.M., Gehrels, G.E., and Dickinson, W.R., 1996, Detrital zircon link between headwaters and terminus of the Upper Triassic Chinle-Dockum paleoriver system: Science, v. 273, p. 97–100.

Royse, F., Jr., 1993, Case of the phantom foredeep: Early Cretaceous in west-central Utah: Geology, v. 21, p. 133–136.

Schumm, S.A., and Thorne, C.R., 1989, Geologic and geomorphic controls on bank erosion, *in* Ports-Michael, A., ed., Proceedings of the 1989 national conference on Hydraulic engineering: New York, American Society of Civil Engineers, Hydraulics Division, p. 106–111.

Stewart, J., and Watts, A.B., 1997, Gravity anomalies and spatial variations of flexural rigidity at mountain ranges: Journal of Geophysical Research, v. 102, p. 5327–5352.

Stewart, J.H., Poole, F.G., and Wilson, R.F., 1972a, Stratigraphy and origin of the Chinle Formation and related Upper Triassic strata in the Colorado Plateau Region: U.S. Geological Survey Professional Paper 690, 336 p.

Stewart, J.H., Poole, F.G., and Wilson, R.F., 1972b, Stratigraphy and origin of the Triassic Moenkopi Formation and related strata in the Colorado Plateau region: U.S. Geological Survey Professional Paper 691, 195 p.

Stokes, W.L., 1944, Morrison Formation and related deposits in and adjacent to the Colorado Plateau: Geological Society of America Bulletin, v. 55, p. 951–992.

Stokes, W.L., 1950, Pediment concept applied to Shinarump and similar conglomerates: Geological Society of America Bulletin, v. 61, p. 91–98.

Swinehart, J.B., Souders, V.L., DeGraw, H.M., and Diffendal, R.F., Jr., 1985, Cenozoic paleogeography of western Nebraska, *in* Flores, R.M., and Kaplan, S.S., eds., Cenozoic paleogeography of west-central United States: Denver, Colorado, Society of Economic Paleontologists and Mineralogists (SEPM) Rocky Mountain Section, Symposium, Volume 3, p. 209–229.

Trimble, D.E., 1980, Cenozoic history of the Great Plains contrasted with that of the southern Rocky Mountains: A synthesis: The Mountain Geologist, v. 17, p. 59–69.

Tschudy, R.H., Tschudy, B.D., and Craig, L.C., 1984, Palynological evaluation of the Cedar Mountain and Burro Canyon Formations, Colorado Plateau: U.S. Geological Survey Professional Paper 1281, p. 24.

Tweto, O., 1979, The Rio Grande rift system in Colorado, *in* Riecker, R.E., ed., Rio Grande rift; tectonics and magmatism: Washington, D.C., American Geophysical Union, p. 33–56.

Walker, J.D., 1988, Permian and Triassic rocks of the Mojave Desert and their implicatons for timing and mechanisms of continental truncation: Tectonics, v. 7, p. 685–709.

Watts, A.B., 1992, The effective elastic thickness of the lithosphere and the evolution of foreland basins: Basin Research, v. 4, p. 169–178.

Winslow, N.S., and Heller, P.L., 1987, Evaluation of unconformities in Upper Jurassic and Lower Cretaceous nonmarine deposits, Bighorn Basin, Wyoming and Montana: Sedimentary Geology, v. 53, p. 181–202.

Witkind, I.J., 1956, Uranium deposits at base of the Shinarump Conglomerate, Monument Valley, Arizona: Washington, U.S. Geological Survey Bulletin, v. 1030-C, p. 99–130.

Yingling, V.L., and Heller, P.L., 1992, Timing and record of foreland sedimentation during the initiation of the Sevier orogenic belt in central Utah: Basin Research, v. 4, p. 279–290.

Zaleha, M.J., Way, J.N., and Suttner, L.J., 2001, Effects of syndepositional faulting and folding on Early Cretaceous rivers and alluvial architecture (Lakota and Cloverly Formations, Wyoming, U.S.A.): Journal of Sedimentary Research, v. 71, p. 880–894.

Manuscript Received by the Society 20 July 2002
Revised Manuscript Received 24 February 2003
Manuscript Accepted 18 March 2003

Printed in the USA