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Synclinal-horst basins: examples from the southern Rio Grande rift and southern transition zone of southwestern New Mexico, USA

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ABSTRACT

In areas of broadly distributed extensional strain, the back-tilted edges of a wider than normal horst block may create a synclinal-horst basin. Three Neogene synclinal-horst basins are described from the southern Rio Grande rift and southern Transition Zone of southwestern New Mexico, USA. The late Miocene–Quaternary Uvas Valley basin developed between two fault blocks that dip 6–8° toward one another. Containing a maximum of 200 m of sediment, the Uvas Valley basin has a nearly symmetrical distribution of sediment thickness and appears to have been hydrologically closed throughout its history. The Miocene Gila Wilderness synclinal-horst basin is bordered on three sides by gently tilted (10°, 15°, 20°) fault blocks. Despite evidence of an axial drainage that may have exited the northern edge of the basin, 200–300 m of sediment accumulated in the basin, probably as a result of high sediment yields from the large, high-relief catchments. The Jornada del Muerto synclinal-horst basin is positioned between the east-tilted Caballo and west-tilted San Andres fault blocks. Despite uplift and probable tilting of the adjacent fault blocks in the latest Oligocene and Miocene time, sediment was transported off the horst and deposited in an adjacent basin to the south. Sediment only began to accumulate in the Jornada del Muerto basin in Pliocene and Quaternary time, when an east-dipping normal fault along the axis of the syncline created a small half graben. Overall, synclinal-horst basins are rare, because horsts wide enough to develop broad synclines are uncommon in extensional terrains. Synclinal-horst basins may be most common along the margins of extensional terrains, where thicker, colder crust results in wider fault spacing.

INTRODUCTION

Many continental rifts are long, narrow tectonic terrains characterized by co-linear or en echelon uplift-basin pairs (Fig. 1a). The most common type of basin is the half graben, which is bordered on only one side by a major normal fault (Rosendahl, 1987). In narrow rifts, individual fault blocks and their complementary basins only interact near the tips of the border faults in regions of strain transfer referred to as accommodation zones (Rosendahl, 1987; Ebinger, 1989; Morley *et al.*, 1990; Mack & Seager, 1995). As it is uplifted, the footwall block commonly tilts away from the border fault. Regardless of whether the driving force is rotational normal faulting (Thompson, 1960; Morton & Black, 1975; Proffett, 1977), reverse drag (Barnett *et al.*, 1987) and/or isostatic effects (Jackson & McKenzie, 1983; Buck, 1988; Wernicke & Axen, 1988), back-tilting of the footwall is contemporaneous with fault activity and constitutes the largest-scale and generally most prominent structural feature of the footwall uplift.

In areas of broadly distributed extensional strain, such as the Basin and Range and southern Rio Grande rift, Tibetan Plateau and Aegean region, faults may be situated such that horst blocks develop, potentially resulting in a structurally and topographically low area within the horst block created by back-tilting of the edges. A sedimentary basin in this setting is herein referred to as a synclinal-horst basin (Fig. 1b). A large-scale version of this appears to be the case for the Tanzanian craton, a gentle sag between the back-tilted eastern and western branches of the East African rift (Gregory, 1921; Holmes, 1945). Occupying the northern part of the sag is Lake Victoria, which developed in the Pleistocene by drainage reversal associated with back-tilting of the western rift margin (Bishop & Trendall, 1967; Scholz *et al.*, 1998).

Presented here are case studies of three synclinal-horst basins in the southern Rio Grande rift and southern Transition Zone of southwestern New Mexico, USA, that illustrate similarities and differences in the history of these

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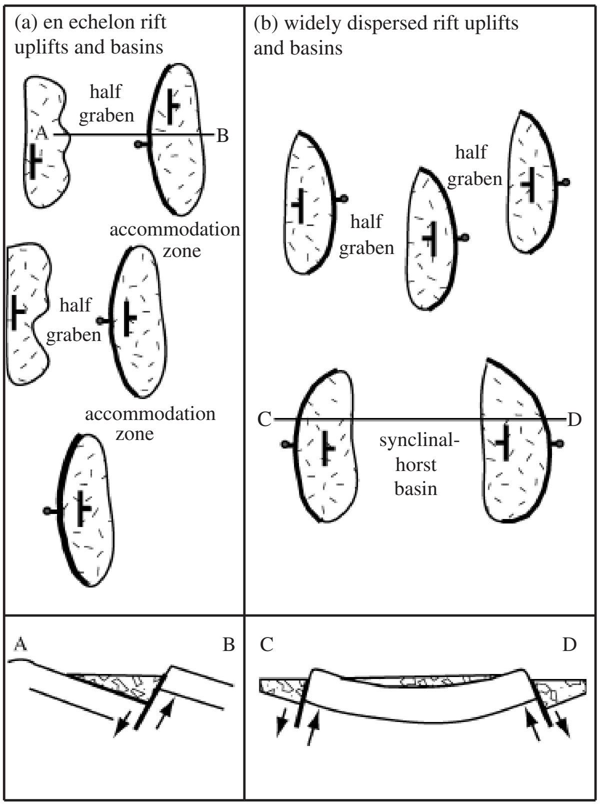

Fig. 1. Schematic geologic maps and cross-sections of (a) en echelon rift uplifts and basins and (b) part of broad rift with widely dispersed fault blocks, two of which dip toward each other and create a synclinal-horst basin.

unusual rift basins, as well as some of the variables that control their development.

GEOLOGIC SETTING

Southwestern New Mexico has experienced latest Paleogene and Neogene volcanism, block faulting and sedimentation resulting from crustal extension in three tectonic provinces, the Basin and Range, southern Rio Grande rift and Transition Zone (Fig. 2). Unlike the northern Rio Grande rift, which consists of a series of right-stepping, en echelon uplift-basin pairs (Chapin & Cather, 1994), strain in the southern Rio Grande rift and Basin and Range in southwestern New Mexico is distributed among dozens of faults in an area roughly 66 000 km^2 (Seager *et al.*, 1982, 1987; Seager, 1995; Hawley *et al.*, 2000). In the part of the southern Rio Grande rift under consideration in this study (right half of Fig. 2), there are more than a dozen major fault blocks that are adjacent to one another in a line perpendicular to the strike of the north-trending faults.

West of the southern Rio Grande rift and north of the Basin and Range in southwestern New Mexico is a Transition Zone that separates those more highly extended terrains from the largely undeformed Colorado Plateau (left half of Fig. 2). Block faulting and concomitant sedimentation in the southern Transition Zone began after the Oligocene and is of smaller magnitude than in the southern Rio Grande rift. Like the southern Rio Grande rift, however,

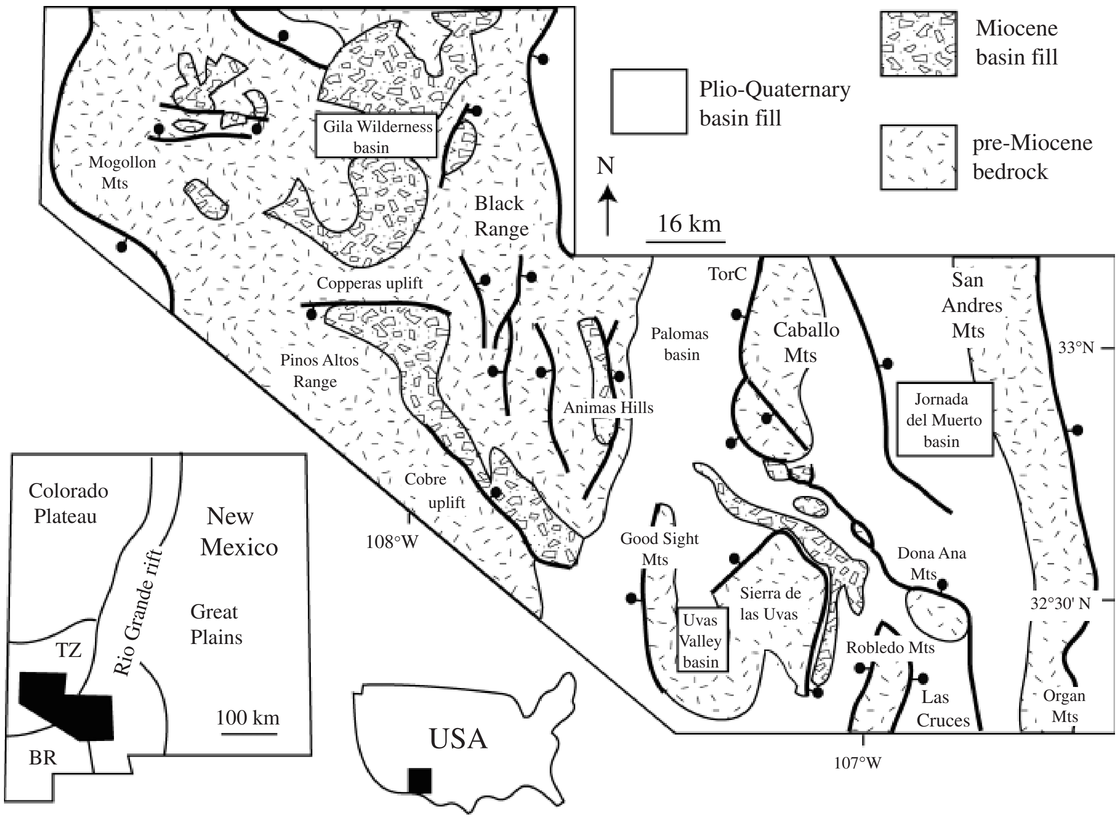

Fig. 2. Generalized geologic map of part of the southern Rio Grande rift and southern Transition Zone, southwestern New Mexico, adapted from Seager *et al.* (1982, 1987). TZ = Transition Zone; BR = Basin and Range.

strain in the southern Transition Zone is accommodated on several faults that result in adjacent, back-tilted fault blocks (Fig. 2; Black Range and Mogollon Mountains).

The stratigraphic record of extension in the southern Rio Grande rift and southern Transition Zone began with latest Eocene or Oligocene volcanism involving eruptions of andesitic and basaltic lava flows, as well as voluminous rhyolitic ignimbrite eruptions from five calderas within the study area and several outside of it (Fig. 3; McIntosh *et al.*, 1991; McMillan *et al.*, 2000). Block faulting and sedimentation were contemporaneous with latest Eocene–early Oligocene volcanism in the southern Rio Grande rift (Mack *et al.*, 1994a). This was followed by a major period of faulting and basin subsidence that began near the Oligocene–Miocene boundary and continued throughout most of the Miocene epoch, resulting in deposition of about 2 km of basin fill of the Hayner Ranch and Rincon Valley Formations (Fig. 3; Mack *et al.*, 1994b). In contrast, large-scale block faulting and basin subsidence in the southern Transition Zone did not begin until the Miocene, resulting in deposition of the Gila Conglomerate (Fig. 3).

A major pulse of deformation in the southern Rio Grande rift began near the Miocene–Pliocene boundary and continues with diminishing intensity to the present day (Seager *et al.*, 1984). During this period new faults developed, many of which inverted parts of Miocene basins. The sedimentary record of this pulse of deformation is the coeval Camp Rice and Palomas Formations, which have a maximum thickness of about 150 m (Fig. 3; Seager *et al.*, 1984; Mack & Seager, 1990). Basin incision and partial backfilling by the ancestral Rio Grande and its tributaries during the past 0.78 Ma deposited a few tens of metres of inset fill and locally exposed all or part of the Camp Rice and Palomas Formations, as well as some older basin-fill strata (Mack *et al.*, 1993, 1998). Faults cutting the Gila Conglomerate indicate post-Miocene tectonism in the southern Transition Zone, but there is little evidence in the study area for large-scale Plio-Pleistocene sedimentation (Fig. 3). Instead, the dominant process appears to have been drainage integration and erosion (Mack, in press).

SYNCLINAL-HORST BASINS IN THE RIO GRANDE RIFT AND SOUTHERN TRANSITION ZONE

Uvas Valley synclinal-horst basin

The Uvas Valley synclinal-horst basin is located between the west-tilted Sierra de las Uvas and the east-tilted Good Sight Mountains (Figs 2 and 4). The basin is late Miocene

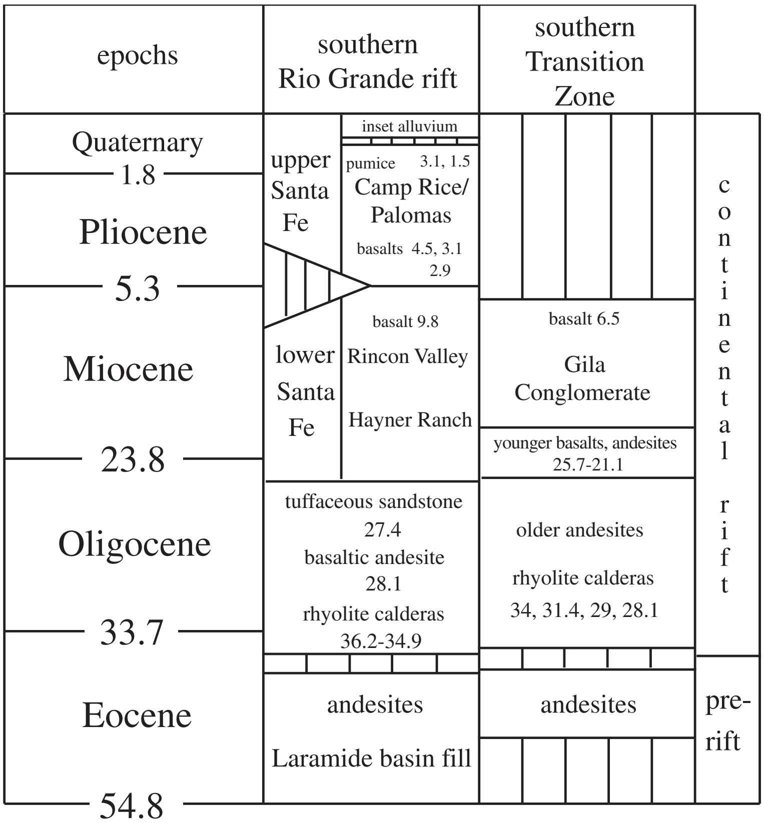

Fig. 3. Late Paleogene and Neogene stratigraphy of the southern Rio Grande rift and southern Transition Zone, southwestern New Mexico. Numbers refer to radioisotopic ages of volcanic rocks in millions of years, including whole-rock basalt and andesite K–Ar dates of 2.9 Ma by Bachman & Mehnert (1978), of 28.1, 9.8, 4.5 and 3.1 Ma by Seager *et al.* (1984), of 25.7–21.1 and 6.5 Ma by Marvin *et al.* (1987), and single-crystal-sanidine ^{40}Ar/^{39}Ar dates of 36.2–34.9, 34, 31.4 and 28.1 Ma by Mcintosh *et al.* (1991), of 27.4 Ma by Boryta & McIntosh (1994) and of 3.1 and 1.5 Ma by Mack *et al.* (1996).

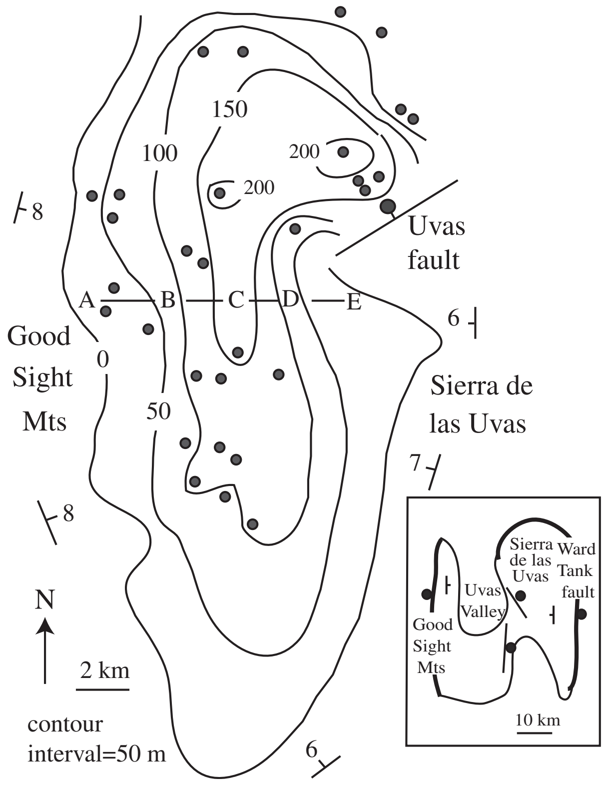

Fig. 4. Isopach map of the upper Miocene Rincon Valley Formation, Pliocene-lower Pleistocene Camp Rice Formation and Quaternary inset alluvium in the Uvas Valley synclinal-horst basin, based on outcrops and well data (black circles) from Clemons (1979). Letters show the location of stratigraphic columns of Fig. 5. The inset shows location of range-bounding faults that create the synclinal-horst basin.

to Quaternary in age and contains up to 200 m of the upper Miocene Rincon Valley Formation and the Pliocene–lower Pleistocene Camp Rice Formation, as well as a few metres to tens of metres of middle Pleistocene to Holocene sediment overlying and locally inset against the Camp Rice Formation (Clemons, 1979). Both the Camp Rice and younger alluvium are exposed along the margins of the basin, whereas sediment thickness in the interior of the basin is constrained by 30 water wells that penetrate the entire basin fill. Logs of the water wells do not distinguish between the Rincon Valley and Camp Rice Formations, although most of the basin fill probably belongs to the latter (Clemons, 1979).

The Uvas Valley synclinal-horst basin is bordered on the east by the east-dipping Ward Tank normal fault that uplifts the Sierra de las Uvas and on the west by a west-dipping normal fault that uplifts the Good Sight Mountains (Figs 2 and 4). The Oligocene volcanic and volcaniclastic bedrock dips gently at 6–8° toward the basin (Clemons, 1979). Located half way between the eastern edge of the basin and the eastern flank of the Sierra de las Uvas are two faults synthetic to the Ward Tank fault. These faults have small displacement and do not affect the overall westward tilt of the range. Mid to late Miocene movement on the Ward Tank fault is indicated by the presence of proximal-fan conglomerates with a Sierra de las Uvas provenance deposited east of the fault (Mack *et al.*, 1994b). A steep gravity gradient and negative anomaly west of the Good Sight Mountains suggest the existence of thick basin fill directly west of the range and imply a long history of movement on the border fault commensurate with inception in Miocene time or earlier (Decker & Smithson, 1975; Daggett & Keller, 1982; Mack *et al.*, 1994a).

Isopachs of the fill of the Uvas Valley basin define a north-trending, nearly symmetrical basin (Fig. 4). The exception to this trend is thick sediment directly adjacent to the Uvas fault, suggesting that the fault was active during the late Miocene, Pliocene and Quaternary time. Exposures of the upper Miocene Rincon Valley Formation a few kilometres northeast of the basin and well cuttings consist of red mudstone deposited in alluvial-flat to lacustrine environments (Fig. 5; Clemons, 1979; Mack *et al.*, 1994b). The Camp Rice Formation exposed along the margin of the basin and in well cuttings consists of alluvial-fan conglomerate and gravel that fine toward the basin axis and a basin-centre facies of interbedded fine gravel, sand and red mud probably deposited in alluvial flat and lacustrine environments (Fig. 5; Clemons, 1979). By the late Pleistocene time, a narrow, north-trending pluvial lake, Lake Goodsight, occupied the northern end of the Uvas Valley basin (Hawley, 1965; Clemons, 1979). This lake, which had a surface area of about 40 km^2, is represented in outcrop by about 5 m of green, ostracod-bearing mud and relict sand and gravel beach ridges (Fig. 5; Clemons, 1979). The basin currently has closed surface drainage and is locally occupied by small, ephemeral lakes. Mapping by Clemons (1979) and well data show that the basin fill onlaps the margins of the basin, because Camp Rice alluvium extends farther toward the surrounding mountains than the Rincon Valley Formation, as does post-Camp Rice, inset alluvium compared to Camp Rice strata (Fig. 5).

Gila Wilderness synclinal-horst basin

The Gila Wilderness synclinal-horst basin was bordered on the east by the Black Range, on the south by the Copperas uplift, and on the west by the Mogollon Mountains, but may have been open to the north (Figs 2 and 6). The basin infilled with Gila Conglomerate, which has a maximum thickness of about 300 m. Throughout the basin the Gila Conglomerate overlies and locally interfingers with basaltic and andesitic lava flows, which yield whole-rock K–Ar radioisotopic ages between 25.7 ± 0.9 and 23.1 ± 0.8 Ma (Marvin *et al.*, 1987). The age of the top of the Gila Conglomerate in the Gila Wilderness basin is not known, but to the southeast, in the Sapillo basin, the Gila Conglomerate is overlain by a basalt dated by the whole-rock K–Ar method at 6.47 ± 0.41 Ma (Marvin *et al.*, 1987).

Prior to development of the Gila Wilderness basin, the area experienced volcanism associated with two calderas, which underwent major eruptions at about 29 and 28.05 ± 0.04 (sanidine $^{40}Ar/^{39}Ar$ date) Ma, as well as post-

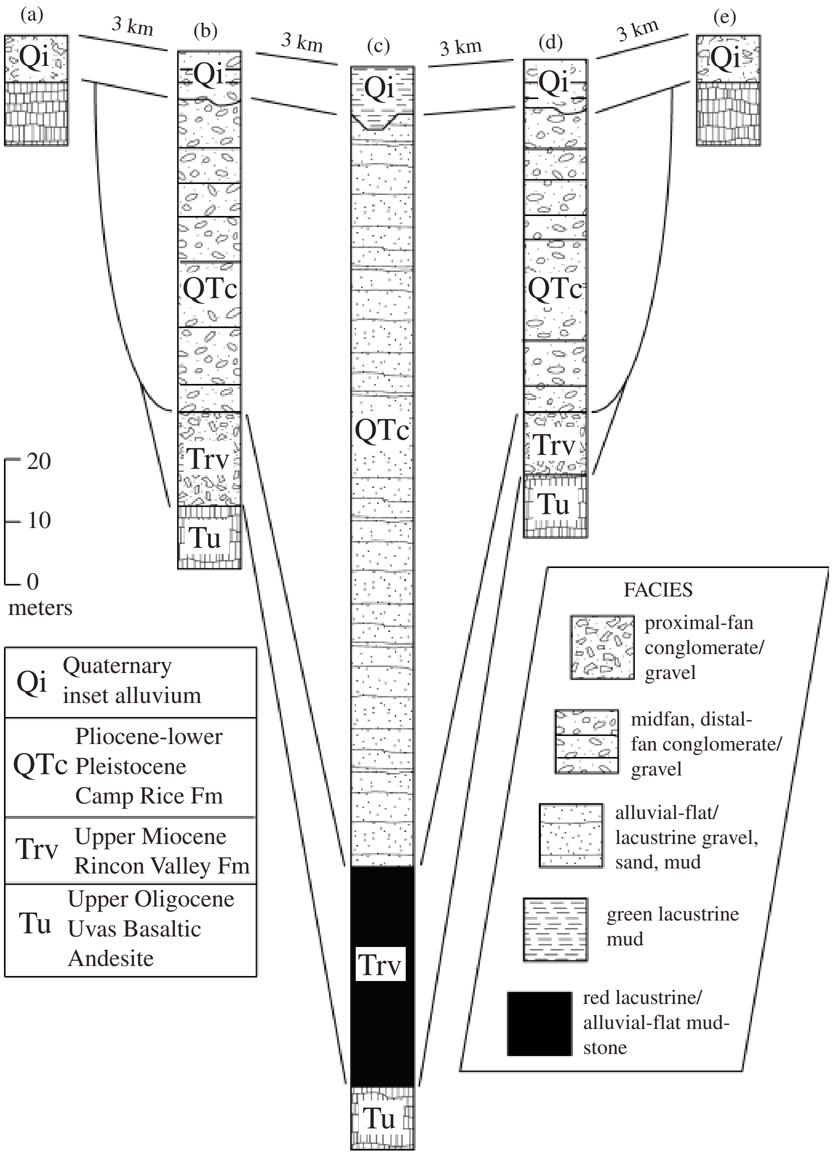

Fig. 5. Facies logs, based on outcrops and well data, across the Uvas Valley synclinal-horst basin. See Fig. 4 for location of lettered sections.

collapse development of flow-banded rhyolite domes and eruptions of rhyolitic tuffs and andesitic lava flows (Ratte *et al.*, 1984; McIntosh *et al.*, 1991). However, much of the topography related to this period of volcanism was levelled by erosion prior to eruption of the younger andesitic and basaltic lava flows (Coney, 1976).

Development of the Gila Wilderness basin in the Miocene appears to have coincided with uplift and tilting toward the basin of the Mogollon Mountains, Black Range and Copperas block (Fig. 6). Evidence for Miocene uplift along the boundary fault of the northern Black Range is provided by the presence in the Winston graben to the east of basin-fill conglomerates interbedded with an andesitic lava flow dated by the whole-rock K–Ar method at 18.3 ± 0.4 Ma (Seager *et al.*, 1984; Harrison, 1994). The southern part of the Black Range appears to have been uplifted at this time as well, based on Miocene conglomerates exposed east of the range that exhibit eastward palaeocurrents and have a Black Range provenance (Fig. 6; Seager *et al.*, 1982). Although complexly faulted, the northern Black Range is generally tilted at a low angle ($\sim 15^{\circ}$) to the west (Coney, 1976).

The timing of uplift of the Copperas fault block is constrained by the depositional history of the Gila Conglomerate in the Sapillo basin (Fig. 6). Here, the Gila Conglomerate overlies a basalt lava flow dated by the whole-rock K–Ar method at 21.2 ± 0.5 Ma and is overlain by the previously discussed 6.47-Ma basalt, making the Gila Conglomerate entirely Miocene in age (Marvin *et al.*, 1987). Coarse, proximal- and mid-fan conglomerates along the northwestern margin of the Sapillo basin have imbrication palaeocurrents directed away from and have clasts derived from the Copperas block, suggesting that it was uplifted during the Miocene time (Wheeler, 1999; Mack, in press). Volcanic bedrock in the Copperas uplift currently dip 10–20° to the north and northeast (Ratte & Gaskill, 1975).

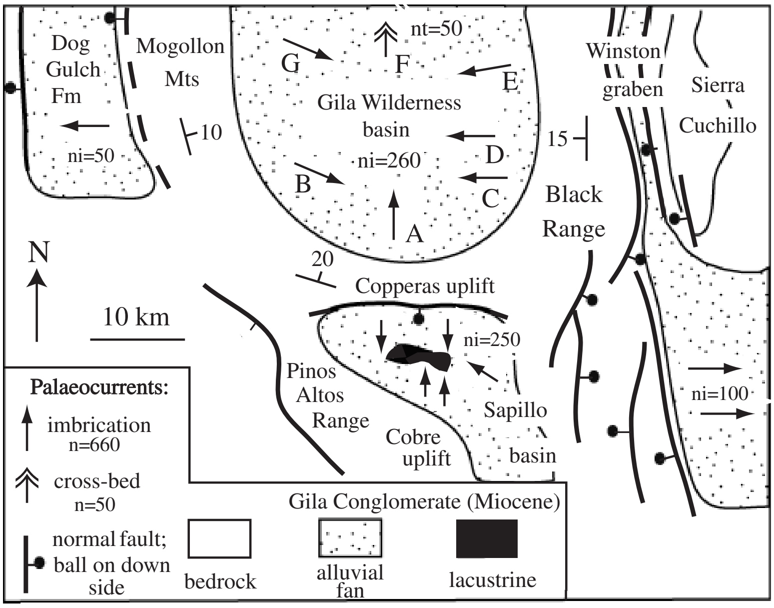

Fig. 6. The Miocene Gila Wilderness synclinal-horst basin and adjacent areas in the southern Transition Zone of southwestern New Mexico. Letters show location of logged sections of Fig. 7. ni = imbrication palaeocurrent data; nt = trough cross-bed palaeocurrent data.

The Mogollon Mountains currently dip gently ($\sim 10^{\circ}$) eastward toward the Gila Wilderness basin, as a result of deformation that is interpreted to have begun during or after eruption of the younger andesites and basalts (Coney, 1976). The fault responsible for initial uplift and tilting of the Mogollon block is, however, not well constrained. Located on the modern footwall block of the Mogollon Mountains are basin-fill conglomerates of the Dog Gulch Formation, whose age is bracketed by basalt flows dated by the whole-rock, K–Ar method at 23.2 ± 0.6 and 15.2 ± 0.5 Ma (Ratte, 1981). Imbrication data from the Dog Gulch Formation indicate westward palaeoflow, introducing the possibility that a fault-bound uplift existed east of the present outcrop of the formation in Miocene time (dashed fault, Fig. 6). Alternatively, the current boundary fault of the Mogollon fault block (solid-line fault, Fig. 6) may have been active in Miocene time, with the Dog Gulch Formation deposited in an intra-uplift basin, as suggested by Coney (1976). Although there are Miocene sedimentary rocks exposed west of the current boundary fault of the Mogollon Mountains, it has not been shown that they were derived from the Mogollon uplift (Houser, 1987, 1994).

The majority of outcrops of the Gila Conglomerate in the Gila Wilderness basin consist of alluvial-fan conglomerates (Fig. 7). Proximal alluvial-fan facies are composed of thick-bedded cobble and boulder conglomerates deposited by debris flows or by current flows in fan channels. Mid-fan to distal-fan conglomerates are thin-bedded, pebble and cobble conglomerates and intercalated coarse sandstones deposited by sheetfloods and hyperconcentrated flows. Imbrication palaeocurrent data from the alluvial-fan conglomerates display drainage toward the basin center (Figs 6 and 7). Also present locally in the north-central part of the basin are pebble and cobble conglomerates and coarse sandstones composed of trough cross-beds in sets 20–100 cm thick. Cross-bed palaeocurrent data indicate palaeoflow to the north, suggesting the existence of an axial drainage with flow depths great enough to create gravelly dunes (Figs 6 and 7). It is not known at this time, however, if the axial drainage existed throughout the entire history of the basin, nor the location of its terminus.

Jornada del Muerto synclinal-horst basin

Located between the east-tilted Caballo Mountains and west-tilted San Andres Mountains is the Jornada del Muerto horst and synclinal-horst basin (Fig. 2). Geologic mapping and well data within the basin, as well as stratigraphic and sedimentologic studies in basins to the south, delineate a three-stage history of uplift and sedimentation in the region.

Initial uplift of the Caballo Mountains fault block took place in latest Oligocene and early Miocene time, corresponding to deposition of the Hayner Ranch Formation (Figs. 3 and 8a; Mack *et al.*, 1994b). Up to 1300 m thick, the Hayner Ranch Formation thickens toward the border fault of the Caballo Mountains. Within a few kilometres of the border fault, the Hayner Ranch Formation consists of proximal- and mid-fan conglomerates whose palaeocurrents are directed away from the Caballo fault block and whose provenance defines an unroofing sequence of Tertiary volcanic rocks (Mack *et al.*, 1994b). By the end of Hayner Ranch deposition, stratigraphic separation on the border fault of the Caballo Mountains is estimated to have been approximately 1600 m (Mack *et al.*, 1994b).

Because there is no basin fill coeval to the Hayner Ranch Formation exposed along the eastern flank of the San Andres Mountains, the exact time of initial uplift of the range is uncertain. However, several lines of evidence suggest that the San Andres Mountains began to rise at about the same time as the Caballo Mountains. First, Precambrian crystalline basement is exposed at the foot of both ranges, implying a similar amount of

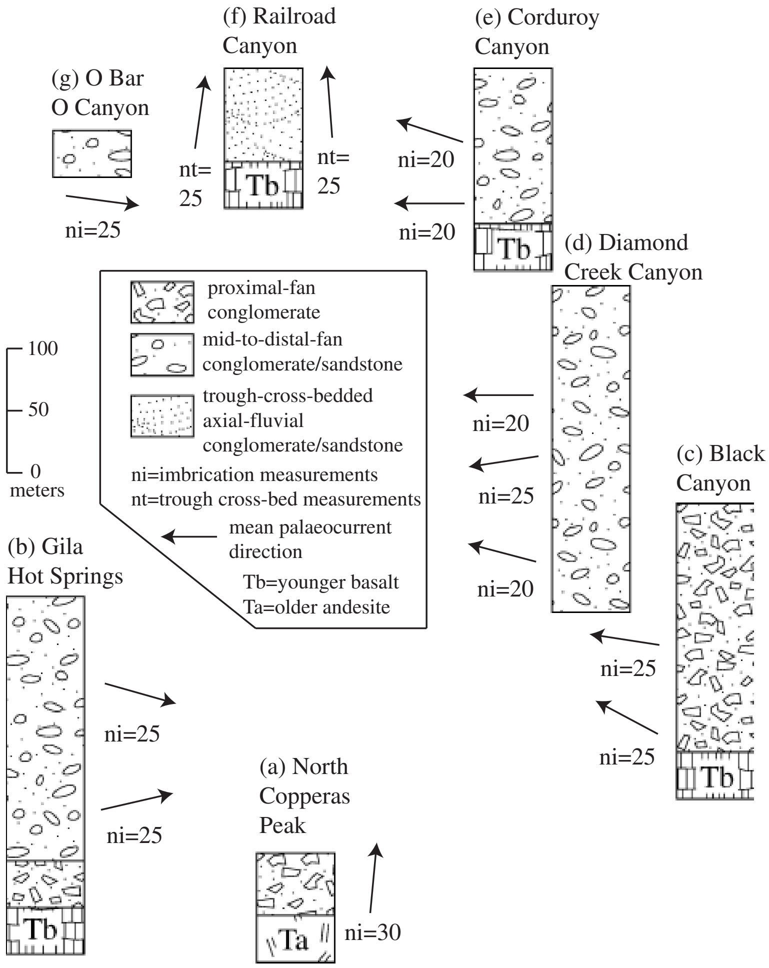

Fig. 7. Logged sections of the Gila Conglomerate in the Gila Wilderness synclinal-horst basin. See Fig. 6 for locations of lettered sections.

stratigraphic separation on the border faults. Second, the degree of tilt ($\sim 20^{\circ}$) of the two ranges is the same. Third, deep exploration wells in the complementary Tularosa basin east of the San Andres Mountains encountered up to 1800 m of basin-fill strata, a thickness comparable to the Hayner Ranch and Rincon Valley Formations combined (Seager, 1981). Thus, if the rates of uplift, back-tilting and sedimentation were comparable between the Caballo and San Andres uplift–basin systems, then the ranges most likely began to rise at about the same time in the latest Oligocene or early Miocene time.

Despite latest Oligocene and early Miocene uplift of the Caballo and San Andres fault blocks, there is no evidence in outcrop or in exploration wells of deposition of Hayner Ranch strata in the Jornada del Muerto basin. It is assumed that sediment eroded from the dip slopes of the Caballo and San Andres Mountains was transported off the Jornada del Muerto synclinal horst into the eastern part of the Hayner Ranch basin, although there are no outcrop-based dispersal data to prove this (Fig. 8a).

During the second stage of development of the Jornada del Muerto basin and surrounding area in the mid to late Miocene time, the Caballo Mountains continued to rise, along with initial uplift of the Sierra de las Uvas, creating two half grabens connected by a playa lake (Fig. 8b; Mack *et al.*, 1994b). An additional 600 m of sediment of the Rincon Valley Formation was deposited adjacent to the Caballo Mountains and upper Carboniferous and Permian rocks were unroofed, indicating an additional 850 m of stratigraphic separation on the border fault of the Caballo fault block by the end of Rincon Valley deposition (Mack *et al.*, 1994b). As was the case with the Hayner Ranch Formation, no Rincon Valley strata are present in the outcrop or in exploration wells in the Jornada del Muerto syncline. In this case, however, sedimentologic data indicate that sediment was transported off the Jornada del Muerto synclinal horst and into the Rincon Valley basin (Fig. 8b). Evidence for such provenance and dispersal are outcrops of distal-fan conglomerates and sandstones of the Rincon Valley Formation exposed at the San Diego Mountain that display southward palaeocurrents and have clasts of Permian and Cretaceous formations exposed only on the present-day dip slopes of the Caballo and San Andres Mountains (Mack *et al.*, 1994b).

By the end of the Miocene or early in the Pliocene, corresponding to the beginning of the final stage in the history of the Jornada del Muerto synclinal-horst basin, a major pulse of deformation occurred in the southern Rio Grande rift (Fig. 8c; Seager *et al.*, 1984; Mack *et al.*, 1994b). Faulting stepped into the former Rincon Valley basin, creating several new fault blocks (Animas Hills, Red Hills, Rincon Hills, San Diego Mountain, Robledo Mountains)

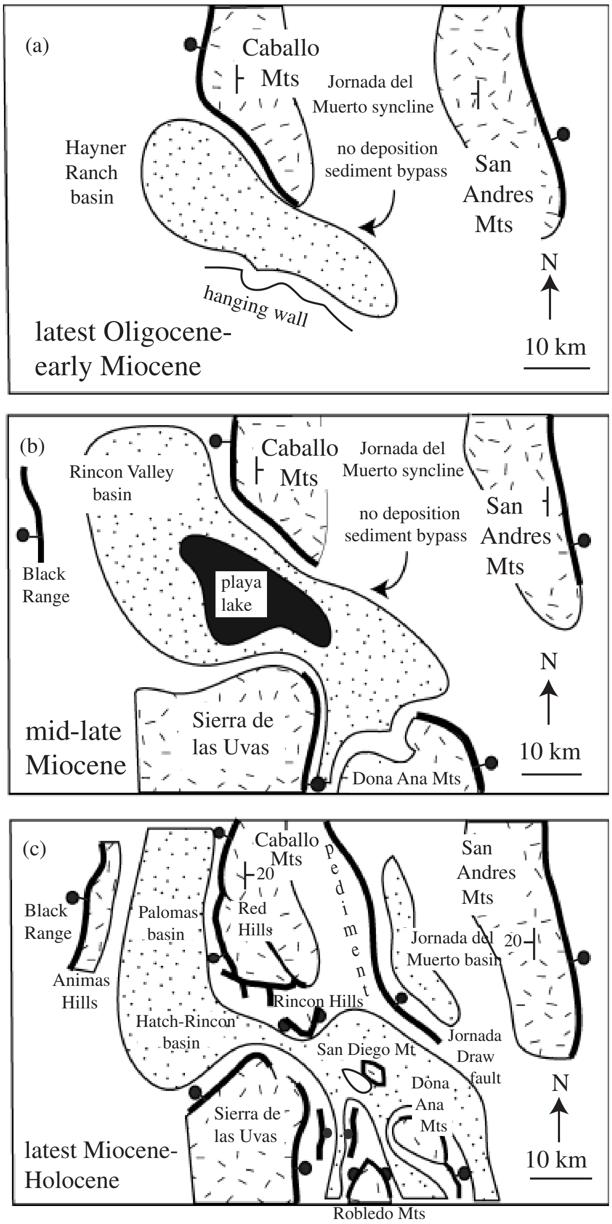

Fig. 8. Late Oligocene to Holocene evolution of the Jornada del Muerto synclinal horst and adjacent areas in the southern Rio Grande rift. Dip angles are not shown for the Caballo Mountains and San Andres Mountains for the latest Oligocene-early Miocene and mid-late Miocene because the values are not known.

and narrower basins (Palomas and Hatch-Rincon basins) (Fig. 8c). At this same time, the Jornada Draw normal fault developed, bisecting the Jornada del Muerto syncline (Seager & Mack, 1995). Sixty-four kilometres long, the Jornada Draw normal fault dips 60° to the east and has a maximum stratigraphic separation near the centre of its trace of about 450 m. The Jornada Draw fault is interpreted to have developed to accommodate uplift and tilting of the Caballo and San Andres Mountains and consequent subsidence of the Jornada del Muerto syncline, because (1) the Jornada Draw fault is parallel to the Caballo and San Andres faults, (2) the Jornada Draw and Caballo faults have similar lengths and latitudes of termination and (3) both the Jornada Draw and Caballo faults have greatest structural relief near the centres of their traces (Seager & Mack, 1995). Once the Jornada Draw fault formed, well data indicate that deposition of 76 m of the Camp Rice/Palomas Formation took place in the half graben east of the fault (Seager *et al.*, 1987). In contrast, to the west of the fault, the Camp Rice/Palomas Formation consists of a thin (<10 m) pediment veneer that truncates the east-dipping bedrock of the Palaeozoic and Cretaceous age. Currently, several small, ephemeral lakes are located directly east of the Jornada Draw fault.

DISCUSSION

Structural constraints on the development of synclinal-horst basins

In regions of active continental extension, the upper crust breaks along normal faults whose length and spacing are proportional to the thickness of the seismogenic layer (Jackson and White, 1989). Although it is generally about 15 km thick, the seismogenic layer locally is more than twice that value, resulting in longer, more widely spaced faults and fault blocks (Jackson & Blenkinsop, 1997). Conversely, in regions of thin, hot crust, the spacing of normal faults is much less than the average value of 15 km (Hayward & Ebinger, 1996). The dominant structural style in regions of continental extension is tilted fault blocks and half grabens bordered on only one side by normal faults (Rosendahl, 1987; Jackson & White, 1989). Full grabens, bordered on both sides by active normal faults, are less common than half grabens, and horsts, bordered by concurrently active, parallel normal faults that dip away from each other, are rare. Like tilted fault blocks and half grabens, the spacing of faults bordering horst blocks is proportional to the thickness of the seismogenic layer, implying that the width of most horsts will be 15 km or less. These observations suggest that synclinal-horst basins should be rare in regions of continental extension, both because horsts themselves are rare and because it should be unusual to find horsts broad enough to allow a syncline of significant width and amplitude to develop between inward-tilted edges of the horst. A phenomenon that may preserve the structural integrity of horst blocks, however, is stress interactions among faults (Hodgkinson *et al.*, 1996; Cowie, 1998). Numerical modelling and empirical data suggest that a stress shadow (region of reduced stress) exists in areas adjacent but transverse to the strike of normal faults. Thus, slip events on the border faults of a horst block may decrease the stress within the horst, preventing it from breaking into smaller faults and sub-basins.

Fault spacing in the southern Rio Grande rift, Basin and Range, and southern Transition Zone of southwestern New Mexico generally follows the observations outlined above. The majority of the basin-bounding normal faults in this region are spaced 10–30 km apart in a line perpendicular to their strikes, with an average spacing of 22.4 km

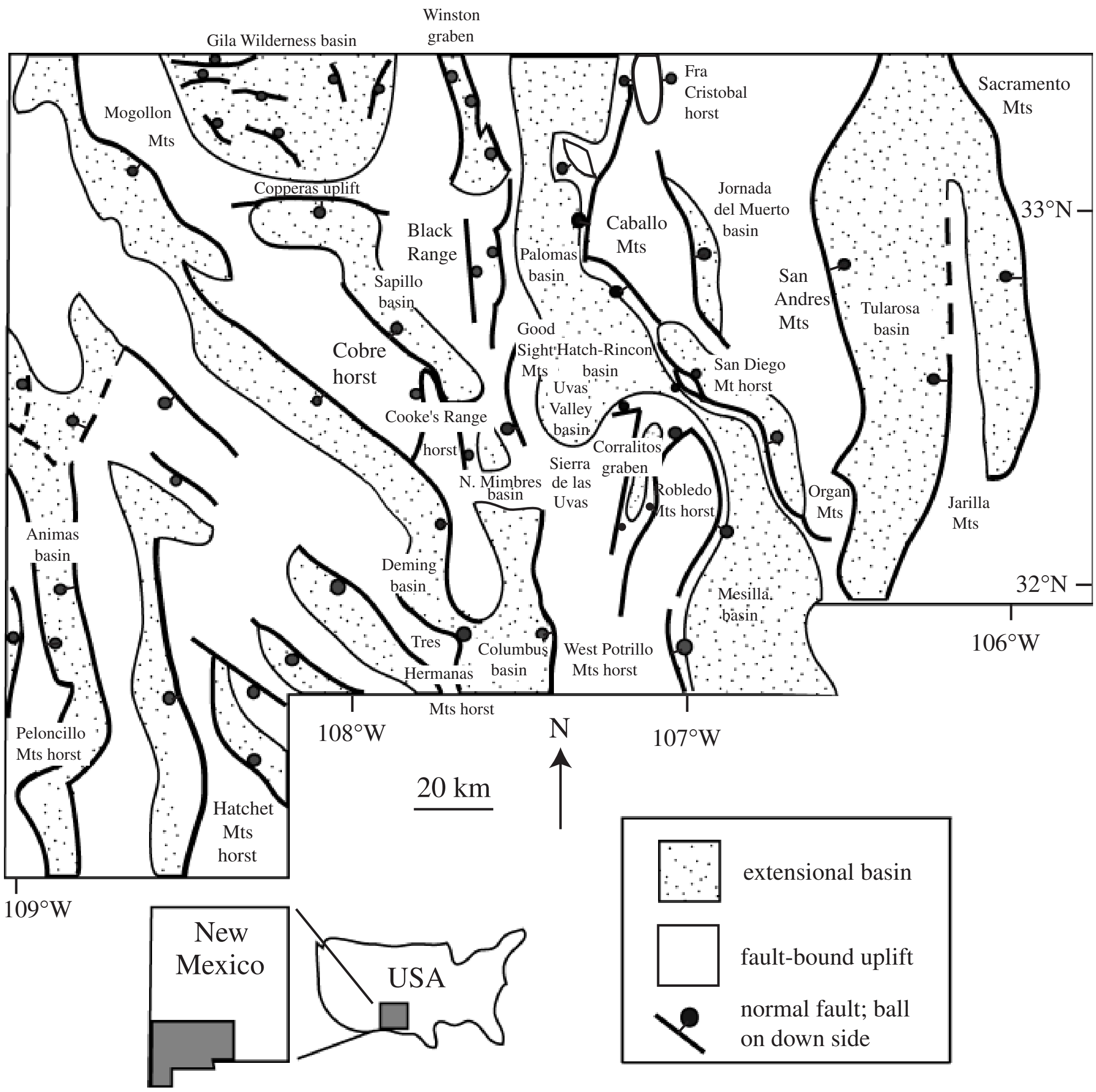

Fig. 9. Map of the southern Rio Grande rift, Basin and Range, and southern Transition Zone in southwestern New Mexico showing major uplifts, basin- and uplift-bounding normal faults and extensional basins, adapted from Seager *et al.* (1982, 1987), Seager (1995) and Hawley *et al.* (2000).

(Figs 9 and 10). Maximum depth of earthquake foci in the Rio Grande rift, which defines the thickness of the seismogenic layer, is close to 20 km (Sanford *et al.*, 1979). Thus, the spacing of the major faults is comparable to the thickness of the seismogenic crust.

The structural style in the southern Rio Grande rift, Basin and Range, and southern Transition Zone also conforms to that in other regions of active crustal extension. Seventeen basins in southwestern New Mexico are half grabens, whereas only eight are full grabens (Fig. 9; full grabens are Animas, Hatch-Rincon, Tularosa, northern Mimbres, Deming, Corralitos, Winston, Columbus basins). In addition to the three synclinal-horst basins that are the subject of this study, there are eight other horsts, which average about 13 km in width (Fig. 9; San Diego Mountain, West Potrillo Mountains, Tres Hermanas Mountains, Cobre uplift, Peloncillo Mountains, Hatchet Mountains, Cooke's Range, Robledo Mountains). The horsts that have synclinal-horst basins exhibit a much wider spacing between border faults than that of the other horsts: 30 km for the Uvas Valley synclinal-horst basin, 60 km for the Jornada del Muerto basin and 90 km for the Gila Wilderness basin.

The origin of two of the anomalously wide horsts with synclinal-horst basins, Jornada del Muerto and Gila Wilderness, may be related to development on thick continental crust along the edges of the region of extension. This appears to be the case in the East African rift, where the eastern and western arms of the rift follow Proterozoic orogenic belts, whereas the crustal sag between the rifts is developed on the thicker Archean Tanzanian craton (Nyblade *et al.*, 1996). Along the eastern margin of the southern Rio Grande rift, the thickness of the continental crust increases steadily toward the Great Plains (Fig. 11). Faults and fault blocks along the eastern margin have an average spacing more than twice that of the average for the entire southern Rio Grande rift, Basin and Range, and southern Transition Zone (cf. Fig. 9; Caballo

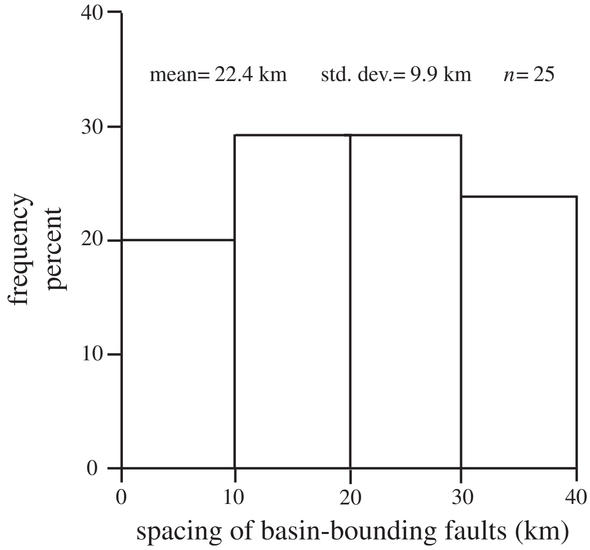

Fig. 10. Histogram of the spacing of major basin-bounding normal faults in the southern Rio Grande rift, Basin and Range, and southern Transition Zone of southwestern New Mexico.

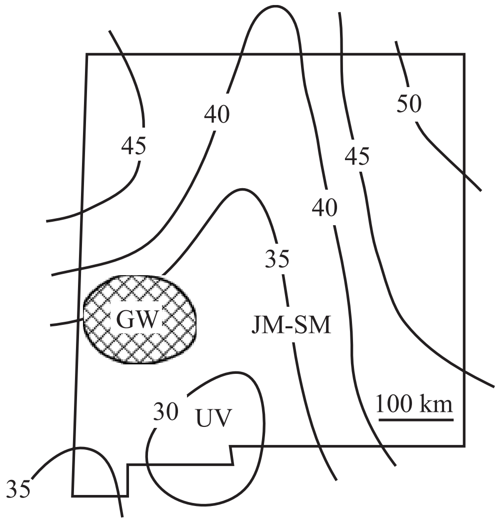

Fig. 11. Crustal thickness in kilometres in New Mexico, taken from Keller (in press). The cross-hatched area corresponds to the location of a pre- to syn-extension, shallow crustal batholith. GW = Gila Wilderness basin; UV = Uvas Valley basin; JM-SM = transect from Caballo Mountains to Sacramento Mountains, including the Jornada del Muerto synclinal horst.

Mountains, San Andres Mountains, Jarilla Mountains, Sacramento Mountains). The eastern margin of the southern Rio Grande rift and the adjacent Great Plains also have experienced little Cenozoic magmatic activity and currently have lower heat flow than the rift axis (Seager & Morgan, 1979). Thus, not only is the crust thicker in this region but it may be colder than the region to the west, both of which may have resulted in a thicker seismogenic layer and more widely spaced faults (Jackson & White, 1989; Jackson & Blenkinsop, 1997).

The Gila Wilderness basin occupies the northwestern margin of the extensional terrain that is transitional to the largely undeformed Colorado Plateau. Here the crust is about 5 km thicker than along the axis of extension (Fig. 11). In addition, the Gila Wilderness basin is superimposed on a large negative gravity anomaly that has been interpreted to represent an upper crustal, granitic batholith about 5 km thick (Fig. 11; Coney, 1976; Schneider & Keller, 1994). This batholith was presumably the source for explosive rhyolitic volcanism associated with calderas whose major eruptions were 29 and 28.1 Ma (McIntosh *et al.*, 1991). Perhaps the granitic batholith made the upper crust more resistant to faulting on a small scale. The existence of a pre-Miocene batholith also introduces the possibility that some component of subsidence of the Miocene Gila Wilderness basin may have been influenced by density increase associated with cooling of the batholith. This is at odds, however, with the observation by Coney (1976) that a period of erosion and not subsidence occurred prior to eruption of the late basalts and andesites that predated deposition of the Gila Conglomerate in the Gila Wilderness basin.

The Uvas Valley synclinal-horst basin is unlike the other two in that it is situated on a region of thin crust associated with high heat flow and evidence of pre- and syn-rift volcanism (Fig. 11; Seager & Morgan, 1979; Seager *et al.*, 1984). This suggests that synclinal-horst basins may develop even along the axis of extensional terrains.

In addition to fault spacing, the degree of back-tilting is also a potential factor in the formation of synclinal-horst basins. Intuitively, deposition in a synclinal-horst basin is related to the accommodation space created by downwarping of the syncline. This suggests that the thickness of the sediment fill should be proportional to the magnitude of back-tilting. This relationship applies to the Uvas Valley and Gila Wilderness basins. The Uvas Valley basin has lower bedrock dips (6–8°) and thinner sediment fill (maximum 200 m) than the Gila Wilderness basin, which has bedrock dips of 10°, 15° and 20° and basin fill up to 300 m thick. The relationship between the degree of back-tilting and sediment thickness does not, however, apply to the Jornada del Muerto basin. Despite a large amount of back-tilting (20°), sediment was not deposited in the Jornada del Muerto synclinal horst until the Jornada Draw fault created a half graben in the centre of the syncline. This may be due to the fact that most of the back-tilting occurred late in the history of the Jornada del Muerto synclinal horst, a factor than cannot be tested with existing data. It is also possible that sediment bypass during the early history of the Jornada del Muerto synclinal horst was related to nontectonic factors, discussed below.

The role of sediment yield and drainage on sedimentation in synclinal-horst basins

In addition to tectonic controls, sedimentation in a synclinal-horst basin may also be influenced by the amount of sediment brought into the basin and by the ability of the drainage system to transport that sediment. The sediment yield from catchments surrounding a synclinal-horst basin is primarily related to the climate, relief and composition of the bedrock. High sediment yield would promote deposition of sediment in a synclinal-horst basin, as long as accommodation space is available, because the drainage system would be overwhelmed with sediment and unable to transport it out of the basin. This may have been the case for the Gila Wilderness basin, which is bordered by large, high-relief mountains that produce high sediment yields. In contrast, it is conceivable that under conditions of very low sediment yield, sediment would not accumulate on a synclinal horst regardless of the creation of accommodation space. Instead, the drainage system would be efficient enough to transport all available sediment out of the syncline. Bypass of sediment through a synclinal horst might also be promoted if the drainage system were connected to an adjacent, topographically lower basin, which appears to have been the case during the early history of the Jornada del Muerto synclinal horst. Finally, synclinal-horst basins with internal drainage, such as the Uvas Valley basin, should accumulate sediment even under conditions of low sediment yield, as long as accommodation space is being created.

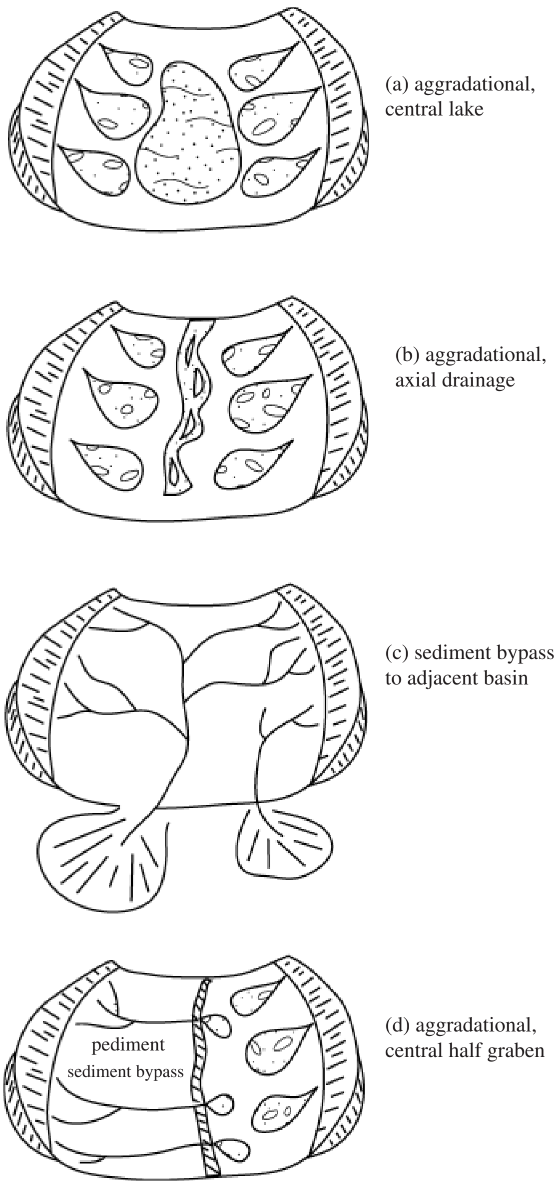

Fig. 12. Summary of the types of synclinal-horst basins encountered in this study. Example A corresponds to the Uvas Valley basin, example B corresponds to the Gila Wilderness basin and examples C and D illustrate the evolution of the Jornada del Muerto basin.

CONCLUSIONS

Synclinal-horst basins may develop in regions of widely distributed extensional strain by back-tilting of the edges of horst blocks. Overall, synclinal-horst basins should be rare compared to half grabens and full grabens, because horst blocks wide enough to develop broad, intra-horst synclines are rare in regions of crustal extension. In the case of the southern Rio Grande rift and southern Transition Zone of southwestern New Mexico, three types of basins developed on horst blocks (Fig. 12). The Uvas Valley basin appears to have been hydrologically closed throughout its history, resulting in ephemeral lakes near the basin centre (Fig. 12a). In contrast, the Gila Wilderness basin, which was larger and had higher relief catchments than the Uvas Valley basin, contained an axial drainage, although its terminus is not known at this time (Fig. 12b). During the early history of the Jornada del Muerto horst, sediment bypassed the synclinal horst and accumulated in an adjacent basin to the south (Fig. 12c). Deposition on the Jornada del Muerto horst did not occur until inception of the Jornada Draw fault, which created a small half graben along the axis of the horst block (Fig. 12d).

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