Contrasting watershed-scale trends in runoff and sediment yield complicate rangeland water resources planning

M. D. Berg[1,*], F. Marcantonio[2], M. A. Allison[3,4], J. McAlister[5], B. P. Wilcox[1], and W. E. Fox[1,5]

[1]{Texas A&M University Department of Ecosystem Science and Management, College Station, Texas}

[2]{Texas A&M University Department of Geology and Geophysics, College Station, Texas}

[3]{The Water Institute of the Gulf, Baton Rouge, Louisiana}

[4]{Tulane University Department of Earth and Environmental Sciences, New Orleans, Louisiana}

[5]{Texas A&M AgriLife Blackland Research & Extension Center, Temple, Texas}

[*]{now at: Save Water Co, Houston, Texas}

Correspondence to: M. D. Berg (mbergtamu@gmail.com)

Abstract

Rangelands cover a large portion of the earth's land surface and are undergoing dramatic landscape changes. At the same time, these ecosystems face increasing expectations to meet growing water supply needs. To address major gaps in our understanding of rangeland hydrologic function, we investigated historical watershed-scale runoff and sediment yield in a dynamic landscape in central Texas, USA. We quantified the relationship between precipitation and runoff and analyzed reservoir sediment cores dated using Cesium-137 and Lead-210 radioisotopes. Local rainfall and streamflow showed no directional trend over a period of 85 years, resulting in a rainfall-runoff ratio that has been resilient to watershed changes. Reservoir sedimentation rates generally were higher before 1963, but have been much lower and very stable since that time. Our findings suggest that (1) rangeland water yields may be stable over long periods despite dramatic landscape changes while (2) these same landscape changes influence sediment yields that impact downstream reservoir storage.

Relying on rangelands to meet water needs demands an understanding of how these dynamic landscapes function and a quantification of the physical processes at work.

1 Introduction

Diverse rangeland ecosystems falling along a grassland–forest continuum cover roughly half of the earth's land surface (Breshears, 2006). Generally precipitation-limited, they are typically used for livestock grazing and harvesting of woody products rather than crop production. But rangelands worldwide face numerous challenges, including (1) conversion to urban development or cultivation; (2) shifting plant cover, such as encroachment by woody plants and invasion by non-native species; and (3) demands for increased production without sacrificing sustainability (Tilman et al., 2002;Van Auken, 2000;Wilcox et al., 2012b).

As growing populations look to these dynamic landscapes to provide critical ecosystem services—including water supply and water storage—their ability to keep pace with these demands is uncertain (Havstad et al., 2007;Jackson et al., 2001). Some of this uncertainty is due to the tremendous variability of runoff and erosion through time and space, which can vary by orders of magnitude even between portions of a single small field (Gaspar et al., 2013;Ritchie et al., 2005). Landscape changes affect these processes further still; and water and sediment yields depend on interactions between climate, vegetation, and local geology. These complex interactions make predictions difficult; and the influence of human activity adds yet another compounding layer of difficulty (Peel, 2009;Boardman, 2006;Vorosmarty and Sahagian, 2000). As a result, major gaps remain in our understanding of rangeland ecosystems. Further interdisciplinary study is imperative to develop a coherent picture of the linkages between hydrological, ecological, and geological processes (Newman, 2006;Wilcox and Thurow, 2006).

Some rangeland investigations have focused on the potential of these landscapes to provide augmented water yields or storage in small reservoirs. Economic and modeling studies have identified vegetation management as a possible means of increasing runoff and streamflow (Griffin and McCarl, 1989;Afinowicz et al., 2005), and government agencies have incorporated these goals into their programs (Texas State Soil and Water Conservation Board, 2005;USDA-NRCS, 2006). Other concerns center on sediment yield, which threatens downstream surface water storage (Bennett et al., 2002;Dunbar et al., 2010). To determine how to respond to these issues and whether related investments are worthwhile, we must gain a better understanding of how rangeland systems function with respect to water resources.

To date, most research has been based on extrapolation of findings from relatively small-scale studies to larger scales or on modeled results. However, because runoff and sediment production are scale-dependent processes, such extrapolation is often unreliable (de Vente and Poesen, 2005;Wilcox et al., 2003). Since they more accurately reveal the true water and sediment yields of watersheds, studies of these processes conducted at the catchment scale are much more relevant to water planning efforts. But whereas catchment-scale data on precipitation and streamflow are somewhat widely available, corresponding sediment data are lacking. Since they serve as archives of historical watershed conditions, the use of reservoir sediments provides one means of filling this data gap and of investigating the impact of human activity (Edwards and Whittington, 2001;Winter et al., 2001). Linking the findings of such investigations with observed changes at the watershed scale will greatly facilitate the development of effective strategies for managing rangeland water resources.

In this study, we investigated the hydrological and sediment transport dynamics of rangeland watersheds. Our main objectives were to (1) quantify long-term trends in precipitation and streamflow using historical data; (2) estimate historical sedimentation rates through radioisotope analysis of reservoir sediment cores; and (3) explore the potential effects of drought conditions on sediment production with historical data. Addressing these objectives not only improves our understanding of rangeland processes but also provides much-needed information on the potential of these landscapes to provide for growing global water needs.

2 Methods

2.1 Study area

As part of a broader study of landscape change and ecosystem function, we examined rangeland processes in the Lampasas Cut Plain of central Texas, USA. This savanna landscape is characterized by low buttes and mesas separated by broad, flat valleys. Local prevailing geology is Cretaceous limestone; soils are loamy and clayey, with occasional sandy loams, and are susceptible to sheet and gully erosion (Allison, 1991;Clower, 1980). The area is drained by the Lampasas River. Streamflow in the upper reaches of the river is runoff-dominated, with localized contributions from springflow (Prcin et al., 2013), and has been recorded at two primary stations (Figure 1). Annual precipitation averages approximately 800 mm, decreasing to the north and west (Figure 2). Winter mean temperature is around 7°C and in summer 27°C.

For the sediment study, we examined eight flood-control reservoirs and their watersheds within the Lampasas River basin. Reservoirs L1, L2, L3, L4, L9, and LX are located in Lampasas County and were constructed between 1958 and 1961. Before impoundment, the parallel watersheds of L1, L2, and L3, contributed to the downstream watershed of LX. Reservoirs M1 and M4, in Mills County, were completed in 1974. Basic attributes of the reservoirs and their watersheds are compiled in Table 1.

Current local land use is predominantly rangeland, and livestock numbers have fluctuated over the last several decades (Figure 3a) while remaining among the highest in the region (Wilcox et al., 2012a). Cropland was widespread early in the 20th century (Figure 3b) but had declined by nearly 80% by 2012 (Berg, M. D., manuscript in review, 2015). Amid this shifting land use, the area has been characterized by large fluctuations in the extent of woody plant cover, due to brush management and regrowth (Figure 3c), and a dramatic increase in the density of farm ponds (Figure 3d) over the last several decades (Berg et al., 2015a).

2.2 Rainfall and runoff trends

To investigate local hydrological trends, we analyzed historical precipitation and streamflow data for the Lampasas River basin. We created a composite record of annual precipitation using a Thiessen polygon approach, centering polygons on available NWS stations (Figure 2). Daily streamflow data were derived from the two USGS stream gage stations downstream from the study watersheds. The lower Youngsport station, with a drainage area of 3,212 km^2, operated between 1924 and 1980; the Kempner station, with a drainage area of 2,119 km^2 has remained active from 1963 to the present.

We performed an automated baseflow separation of streamflow data from each station (Arnold and Allen, 1999). This digital filter approach is objective and reproducible and partitions annual baseflow and stormflow with high efficiency (Arnold et al., 1995)—enabling these components to be interpreted in light of changing landscape conditions.

Using the precipitation (1924—2010) and two streamflow datasets (1924—1980; 1963—2010), we applied a nonparametric Mann-Kendall trend test (Lettenmaier et al., 1994) to detect directional changes in precipitation, total streamflow, baseflow, and stormflow. We performed two-tailed statistical tests for significance, with $\alpha = 0.10$.

2.3 Reservoir sedimentation rates

To shed light on sediment transport processes, we extracted cores from each of the eight reservoirs and analyzed sediments using Cesium-137 (^{137}Cs) and Lead-210 (^{210}Pb) tracers. ^{137}Cs is present in the environment as a result of atomic weapons testing and accidental emissions. ^{210}Pb occurs naturally. Both can be used to estimate sedimentation rates and interpret transport history in a variety of environments (Walling et al., 2003;Ritchie and McHenry, 1990;Appleby and Oldfield, 1978). Coring sites were selected by locating the thickest sediment deposits through exploratory hydroacoustic surveys (U.S. Army Corps of Engineers, 2013, 1989;Dunbar et al., 2002). In each reservoir, we extracted sediment cores at identified sites near the dam structure, from locations corresponding to the pre-impoundment floodplain (Figure 4). Taking cores from these areas reduces the likelihood of capturing mixed profiles, which skew analysis (Sanchez-Cabeza and Ruiz-Fernández, 2012). It also ensures the collection of fine sediments, to which the radioisotopes preferentially adsorb (Bennett et al., 2002). We extracted cores using a portable vibracoring system suspended from a floating platform. This method captures unconsolidated, saturated sediments with minimal disturbance and compaction (Lanesky et al., 1979). The cores were collected with an aluminum pipe lowered to the point of refusal, penetrating the pre-impoundment surface. Retrieved cores were sealed and transported upright to cold storage (~5°C).

We sectioned each core vertically in 3-cm intervals, drying each section for analysis according to IAEA (2003) protocols. A subsample of each core section was ground to homogenize its contents, sealed in a 50 mm x 9 mm Petri dish, and allowed to ingrow for at least 21 days so that ^{210}Pb supported levels reached equilibrium. Counts for ^{210}Pb and ^{137}Cs were performed according to Hanna et al. (2014) using a Canberra low-energy germanium gamma spectrometer. Radioisotope activity was indicated by photopeaks at 46 keV (total ^{210}Pb) and 661.6 keV (^{137}Cs). Excess ^{210}Pb was calculated by subtracting the supported activity of the ^{226}Ra parent—obtained by averaging the 295, 351.9, and 609.3 keV peaks of the ^{214}Pb and ^{214}Bi daughter products—from total measured ^{210}Pb activity at the 46 keV peak. Activity measurements were validated with IAEA-300 standard reference material.

To determine historical linear sedimentation rates, we used as a chronological marker the depth of peak ^{137}Cs activity (corresponding to the 1963 peak in global atmospheric fallout) (Ritchie et al., 1973). We calculated average linear sedimentation rates for the post-1963 period by dividing this depth by the time elapsed between 1963 and the coring date for each reservoir; we calculated the pre-1963 rates by dividing the depth of sediment below the activity peak by the time elapsed between reservoir impoundment and 1963.

To complement ^{137}Cs analysis, we used excess ^{210}Pb activities to calculate the linear sedimentation rate for each core (Krishnaswamy et al., 1971;Bierman et al., 1998). We also searched for changing deposition rates within each core, as plots of the natural log of excess ^{210}Pb versus depth indicate stable sedimentation rates over time when R^2 approaches 1.0.

Finally, we obtained historical annual Palmer Modified Drought Index (PMDI) data for the region to identify potential climatic drivers of sedimentation during different periods. We plotted PMDI and annual peak flows (from USGS data) between 1924 and 2010, identifying episodes conducive to increased sediment production (in particular, a wet year or years following a period of intense drought).

3 Results

3.1 Rainfall and runoff trends

Despite a great deal of interannual variability, there was no directional change in local precipitation 1924—1980 ($p = 0.90$) or 1963—2010 ($p = 0.22$), which has remained near a long-term average of 800 mm (Figure 5a). The same is true of total streamflow (1924—1980: $p = 0.98$, 1963—2010: $p = 0.34$), which has averaged between 60 and 70 mm (Figure 5b). As a result, the rainfall–runoff ratio, the proportion of rainfall leaving a watershed as streamflow, also remained unchanged, at approximately 8% (1924—1980: $p = 0.90$, 1963—2010: $p = 0.45$). Moreover, neither baseflow nor stormflow exhibited a directional change over either period of record. However, baseflow as a proportion of total streamflow did increase 1924—1980 ($p = 0.02$) despite minimal change in overall flow—almost doubling its contribution (Figure 5c).

3.2 Reservoir sedimentation rates

Sediment core profiles varied widely in depth between reservoirs—from less than 3 cm in LX to 162 cm in L1 (Figure 6). Activity peaks of ^{137}Cs supported the analysis of pre-1963 sedimentation rates for reservoirs L1, L2, L3, and L9. Overall, linear sedimentation rates were higher before 1963 (Table 2; Figure 7). Except in the case of L3, sediment deposition has slowed since 1963—by 54% in L1, 76% in L2, and 84% in L9. In reservoir L3, it increased by 49% after 1963. Reservoir L1 exhibited the highest sedimentation rate both before and after 1963. However, when normalized by catchment area, sedimentation rates varied much more widely. That in L9 was by far the highest—surpassing the next highest reservoir by nearly 1400% for the pre-1963 period and by 423% for the post-1963 period.

Cores from L4, LX, M1, and M4 did not display a ^{137}Cs peak. For these cores, sedimentation was assumed to be post-1963 and was estimated by dividing sediment depth by time since impoundment. For cores L4 and M4, which did not capture the entire sediment profile, actual rates likely are higher than those calculated.

Cores from reservoirs LX and M1 showed vertical mixing that prohibited ^{210}Pb analysis. However, remaining cores displayed high correlation between ^{210}Pb activities and depth, indicating linear sedimentation rates have remained quite stable over time (Table 2). ^{210}Pb-based estimates generally resembled those based on ^{137}Cs activities. In addition, rates calculated from ^{210}Pb activities were similar to the post-1963 rates based on ^{137}Cs activities ($p = 0.84$), suggesting good agreement between the two methods for the period since 1963.

Chronological data revealed periods of drought of varying intensity and occasional years of very high streamflow (Figure 8). The historic 1950s drought was longer and more severe than any other over the last century; it was followed by periods of very high flow in 1957 and 1960. Comparable high flows in 1965 occurred in the middle of a multi-year drought, and the severe drought beginning in 2006 featured occasional elevated peak flows. In 1992, very high flows occurred during a prolonged wet period.

4 Discussion

4.1 Rainfall and runoff trends

Given the varying trends in precipitation and streamflow observed in many regions (Lins and Slack, 1999;Andreadis and Lettenmaier, 2006), the dynamic hydrological stability in our study area is surprising. At the same time, such consistency sheds light on the effects of watershed changes on local water budgets. Studies at small spatial scales frequently indicate that landscape changes have important water resource impacts, with the specific response depending on the relative importance of evapotranspiration, recharge, and runoff (Foley et al., 2005;Kim and Jackson, 2012). Such changes affect local water budgets and influence water yields (Petersen and Stringham, 2008;Huxman et al., 2005;Farley et al., 2005). However, complicated feedbacks make effects at larger scales highly uncertain and often overwhelmed by climatic and physical characteristics (Peel, 2009;Wilcox et al., 2006;Kuhn et al., 2007). Our rainfall–runoff ratio of 8% is essentially identical to early estimates of 7% for the area (Tanner, 1937). The lack of a directional trend in streamflows suggests that this region, like many semiarid landscapes dominated by surface runoff, is largely hydrologically insensitive to shifting watershed characteristics (Wilcox, 2002). Perceived impacts due to changing rooting depths, longer growing seasons for evergreen woody plant species, and assumptions of very high shrub transpiration capacities are not borne out. Changes in land use and land cover—and even the impoundment of small reservoirs—have had negligible impacts on streamflow. These results confirm and add new insight to other research showing that woody plants in this region are shallow-rooted and do not rely on deeper, perennial water sources (Heilman, 2009;Schwartz et al., 2013;Schwinning, 2008).

It is still not understood why baseflow showed a proportional increase 1924—1980. In some landscapes, improving range conditions have led to increased infiltration (Wilcox and Huang, 2010). However, local livestock numbers have remained high, and karst features are limited—unlike other regions where baseflow increases have been attributed to rangeland recovery. It is possible that infiltration from local impoundments has added to baseflows. Despite minimal effects on total streamflow, even small dams can create localized groundwater recharge (Graf, 1999;Smith et al., 2002), and Lampasas River tributaries are characterized by a high degree of connectivity between surface water and local aquifers (Mills and Rawson, 1965).

Perennial flow in this part of the Lampasas River is maintained by isolated springs fed by an aquifer extending beyond the basin (Mills and Rawson, 1965). As a result, the effective catchment of the river is larger than it appears, and springflow contributions complicate the interpretation of streamflows. At the same time, it is clear that the fundamental relationship between rainfall and streamflow has not changed over more than 85 years—suggesting that the Lampasas River is hydrologically resilient in the face of changing land use and land cover.

4.2 Reservoir sedimentation rates

Because sediment deposition affects reservoir storage and flood detention, understanding sedimentation rates over time is critical to managing rangeland water resources. Though questions do remain regarding the opposing trend in reservoir L3, changes in rates make it clear that sedimentation was more rapid before 1963. The period since that time has been characterized by stable and lower yields. But what explains the higher rates seen during the earlier period? Additional historical landscape data may offer a key interpretive lens.

Livestock can be powerful instruments of landscape change, both directly (trampling soils) and indirectly (disturbing protective vegetation). When grazing is prolonged or intense, sediment yield can be great (Trimble and Mendel, 1995). The high animal densities in this area around the time of reservoir impoundment doubtless contributed to erosion (Figure 3a).

Crop production also can result in accelerated erosion by damaging soil structure and depleting organic matter (Quine et al., 1999). Cropland is a major source of sediment in many landscapes (Foster and Lees, 1999;Blake et al., 2012). In our study area, cropland acreage has declined dramatically since the 1930s (Figure 3b). Further, nationwide improvements in soil conservation have reduced sediment yield from many agricultural lands (Knox, 2001).

While woody plant encroachment influences soil loss, removing undesirable shrubs and trees also elevates short-term sediment yields (Porto et al., 2009). Since the time of initial settlement, woody plant management has resulted in major land cover changes (Figure 3c). Most early removal was done manually, and the first mechanical control methods were very destructive, leading to high erosion rates (Hamilton and Hanselka, 2004). In recent decades, however, brush removal has declined with shifting landowner priorities (Sorice et al., 2014).

Changes in precipitation frequency, duration, or intensity also affect sediment transport (Xie et al., 2002;Allen et al., 2011). Similarly, drought is an important driver of sediment dynamics in many rangelands. Extended dry periods can cause long-term shifts in plant cover, leading to sediment pulses when rains return (Allen and Breshears, 1998;Nearing et al., 2007). The Lampasas River experienced very high flows in 1957, 1960, 1965, and 1992, and some of these were associated in time with severe droughts (Figure 8). Just before the impoundment of most of the reservoirs we examined, the region was in the grip of drought conditions unmatched since European settlement (Bradley and Malstaff, 2004). Our sediment records cover only the end of this drought but show pre-1963 deposition 220–630% faster than subsequent rates. However, any direct effects of the 1957 drought-breaking floods would not be found in the sediments of the reservoirs, which were impounded beginning in 1958. Interestingly, we also did not find spikes in sedimentation associated with high flows or droughts later in the study period. The apparent low importance of drought and floods in sediment delivery in these watersheds is surprising.

Together, these factors have acted over multiple temporal and spatial scales to influence sediment yields in the study area. Yet because there is no clear link between contemporary land use, land cover, and sedimentation rates, it is possible that another process has reduced sediment yields.

4.3 Sediment storage

To truly understand the local sediment processes at work, it is important to understand what our findings actually show. Sedimentation rates are poor indicators of in-field soil erosion and redistribution (Nearing et al., 2000;Ritchie et al., 2009); what they do reflect is more closely related to net watershed sediment yield. Sediment yield is buffered by internal storage. Especially at larger scales, watersheds can have a great deal of internal storage, so that very little eroded soil actually leaves the watershed, even in the presence of extreme erosion (Bennett et al., 2005;Porto et al., 2011).

In this study area, the increasing density of farm ponds (Figure 3d) represents a key potential sink for watershed sediments. These ponds – usually < 0.3 ha when full – retain material that otherwise would be transported downstream, reducing sediment yields. Because of their smaller contributing watersheds, ponds have high trap efficiencies, magnifying their effects (Brainard and Fairchild, 2012). Indeed, impoundments may be the single greatest anthropogenic modifier of sediment transport; globally, most sedimentation now takes place in aquatic settings and will be retained therein for long periods (Renwick et al., 2005;Verstraeten et al., 2006).

In addition to this storage of eroded sediments in local ponds, a vast amount of sediment from past erosion likely remains on the landscape (Beach, 1994;Meade, 1982). The initial decades after European settlement in this area saw intensive cultivation and very high livestock densities (Jordan-Bychkov et al., 1984;Wilcox et al., 2012a). This destructive combination remained in place for nearly a century in the Lampasas Cut Plain. By the 1930s, many rangelands were already seriously degraded (Mitchell, 2000;Bentley, 1898;Box, 1967). While the methods we used do not allow us to determine whether reservoir sediments result from contemporary erosion or are a legacy of earlier land use, stabilizing sediment yields and observations of local gully erosion suggest that deposits from prior erosion continue to be a source of sediment (Bartley et al., 2007;Mukundan et al., 2011;Phillips, 2003).

The lack of sediments in LX appears to lend support to the importance of internal deposits. This reservoir's watershed is comparable in size to those of L2, L3, and M4, yet sedimentation rates were only 3%–14% of those in the other reservoirs. When L1, L2, and L3 were impounded, the effective catchment area of LX decreased by 86%. Without the historical streamflows and sediment loads from those tributaries, deposits are no longer mobilized and transported downstream.

Given this complexity, we suggest that radioisotope tracers have great potential to elucidate the dynamics of rangeland systems, particularly as their use evolves from primarily research applications to use as a management and decision-support tool (Mukundan et al., 2012). Further strides can be made in understanding rangeland processes by (1) incorporating historical climate, land use, and land cover information to interpret sediment data (Venteris et al., 2004;Boardman, 2006) and (2) including sediment surveys of the farm ponds that are much smaller yet far more abundant than the reservoirs we examined (Downing et al., 2006).

5 Conclusion

We examined long-term trends in rainfall, runoff, and sediment yield in rangeland watersheds with a dynamic land use history. Over more than 85 years, neither precipitation nor streamflow showed any directional trend, suggesting a lack of hydrological sensitivity to landscape change. This raises doubts over efforts to increase runoff by directing land cover changes. Reservoir sedimentation rates generally were higher before 1963, and then stabilized at a lower level over the 50 years since 1963. We believe that this decline in sediment yield is related to long-term landscape changes and an increase in internal storage. As a result, future changes in land use or sediment storage may impact downstream reservoir capacity. These findings challenge simplistic assumptions about streamflow and sediment yield in dynamic rangelands. Determining the role of these landscapes in meeting growing water resource demands requires a creative approach. Integrating multiple techniques with historical information enables a more complete understanding of rangeland processes and holds the key to informed water planning.

Data availability

Streamflow data are available at the USGS National Water Information System. Stream gages: 08103800 (Kempner) and 08104000 (Youngsport). Drought data are available at the NOAA National Climate Data Center. Texas Climate Division: CD 3 (North Central) and CD 6 (Edwards Plateau).

Acknowledgements

Dan Duncan, Andrea Hanna, and Diana di Leonardo performed activity counts of sediment samples. This work was supported by USDA-NIFA Managed Ecosystems grant 2011-68002-30015, USDA-NIFA National Needs Program grant 2009-38420-05631, NSF-CNH grant 413900, and a Tom Slick Graduate Research Fellowship from the Texas A&M University College of Agriculture and Life Sciences.

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Table 1. Sediment study reservoirs and watershed characteristics.

Reservoir	Primary Inflow	Surface Area (km^2)	Watershed Area (km^2)	Year Impounded	Year Cored	Min. Elev. (m)	Max. Elev. (m)
L1	Donalson Creek	0.20	50.9	1959	2010	367	500
L2	Pitt Creek	0.18	23.2	1959	2010	362	458
L3	Espy Branch	0.11	27.5	1958	2010	355	459
L4	Pillar Bluff Creek	0.07	41.2	1960	2012	345	467
L9	Cemetery Creek	0.02	1.2	1960	2012	322	363
LX	Bean Creek	0.20	23.1	1961	2012	338	420
M1	Middle Bennett Creek	0.14	34.6	1974	2012	422	536
M4	Mustang Creek	0.15	28.0	1974	2012	432	534

Table 2. Linear sedimentation rates derived from radioisotope activities.

Core	^{137}Cs				^{210}Pb		
	Pre-1963		Post-1963		Core mean		R^2
	cm y^{-1}	cm y^{-1} km^{-2}	cm y^{-1}	cm y^{-1} km^{-2}	cm y^{-1}	cm y^{-1} km^{-2}	ln dpm g^{-1}
L1	6.4	0.13	2.9	0.06	3.1	0.06	0.90
L2	3.4	0.15	0.8	0.03	0.9	0.04	0.97
L3	1.4	0.05	2.1	0.08	1.3	0.04	0.96
L4	[a]	[a]	0.5[b]	0.01[b]	1.2	0.03	0.93
L9	2.5	2.02	0.4	0.32	0.4	0.19	0.94
LX	[a]	[a]	0.1	< 0.01	[c]	[c]	[c]
M1	[a]	[a]	1.5	0.04	[c]	[c]	[c]
M4	[a]	[a]	0.4[b]	0.01[b]	0.8	0.01	1.00

[a] Core did not display a ^{137}Cs peak, and rates were calculated using the time elapsed since impoundment.

[b] Core did not capture the pre-impoundment surface and likely underestimates true values.

[c] Core showed significant vertical mixing, preventing calculation of sedimentation rate.

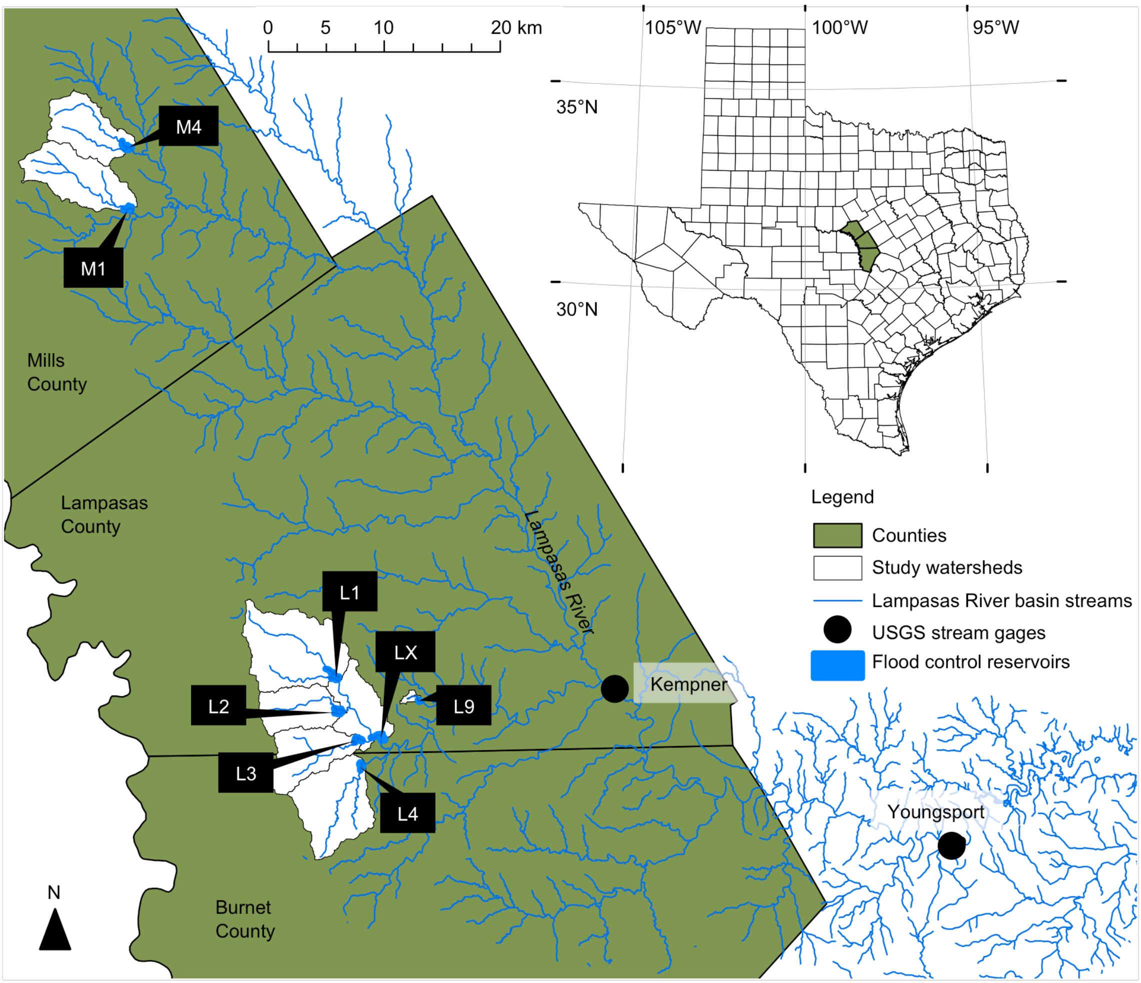

Figure 1. Study area in Texas, USA. Each study watershed encloses a flood control reservoir from which sediment cores were collected. All watersheds contribute flow to the Lampasas River.

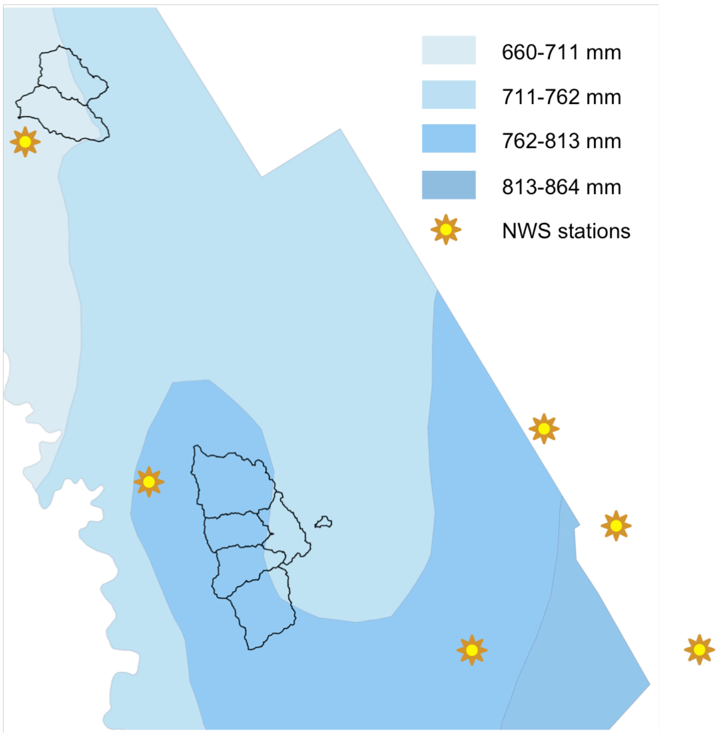

Figure 2. Average annual precipitation gradient and location of National Weather Service (NWS) stations used to construct historical precipitation record.

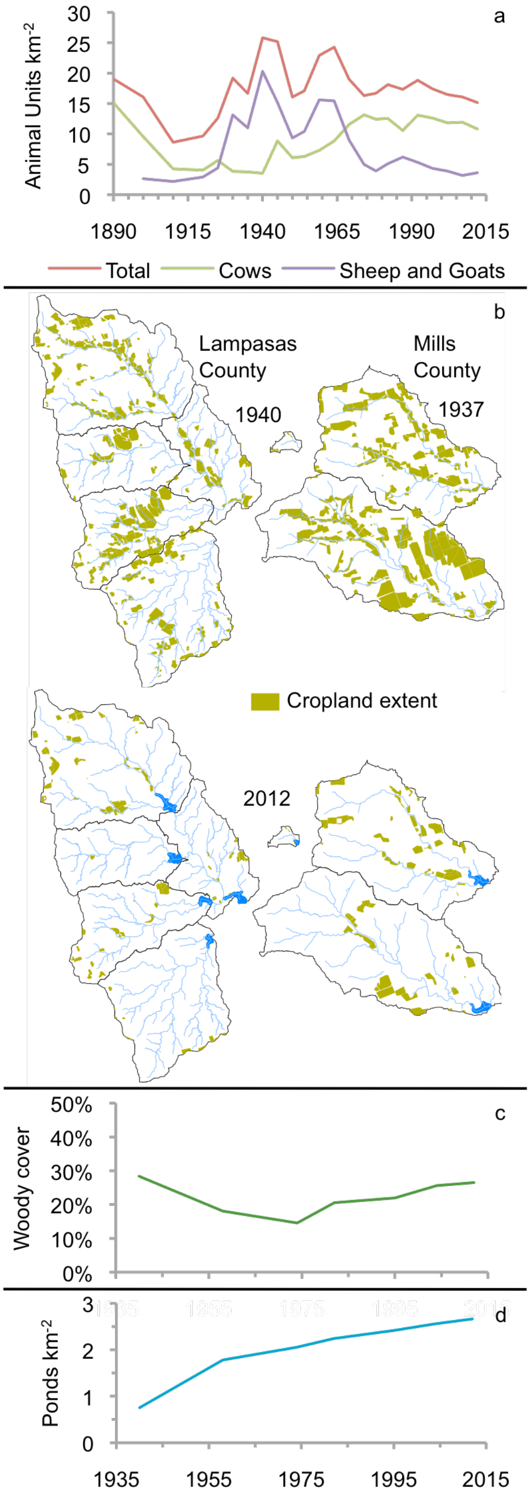

Figure 3. Historical landscape changes in the study area. (a) Livestock numbers in the Lampasas Cut Plain. Recreated from Wilcox et al. (2012a). (b) Extent of active cropland in 1937-40 and 2012 (Berg, M. D., manuscript in review, 2015). (c) Historical extent of woody plant cover in the study watersheds (Berg et al., 2015b). (d) Pond density over time in the study watersheds (Berg et al., 2015a).

Figure 4. Reservoir sediment coring apparatus (top) and representative sediment profile (bottom).

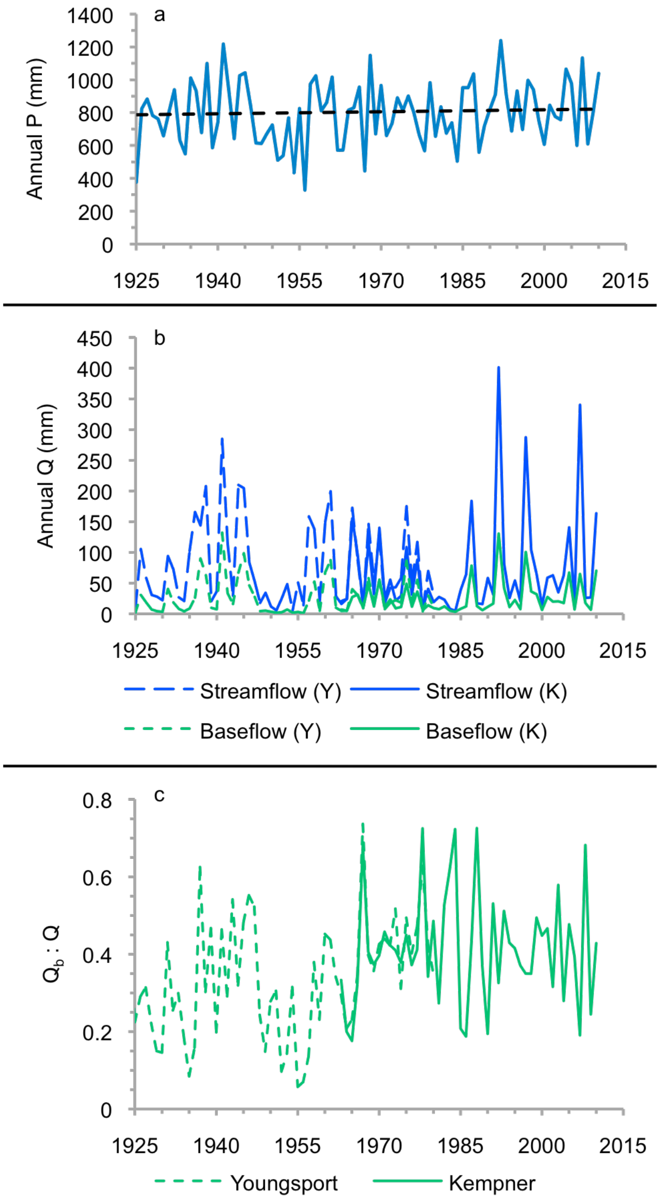

Figure 5. Precipitation and streamflow trends of the Lampasas River basin. (a) Precipitation showed no directional trend. (b) Streamflow showed no directional trend at either the Youngsport (Y) or Kempner (K) station, despite being highly variable. (c) Baseflow as a proportion of total streamflow displayed an upward trend over the first portion of the study period.

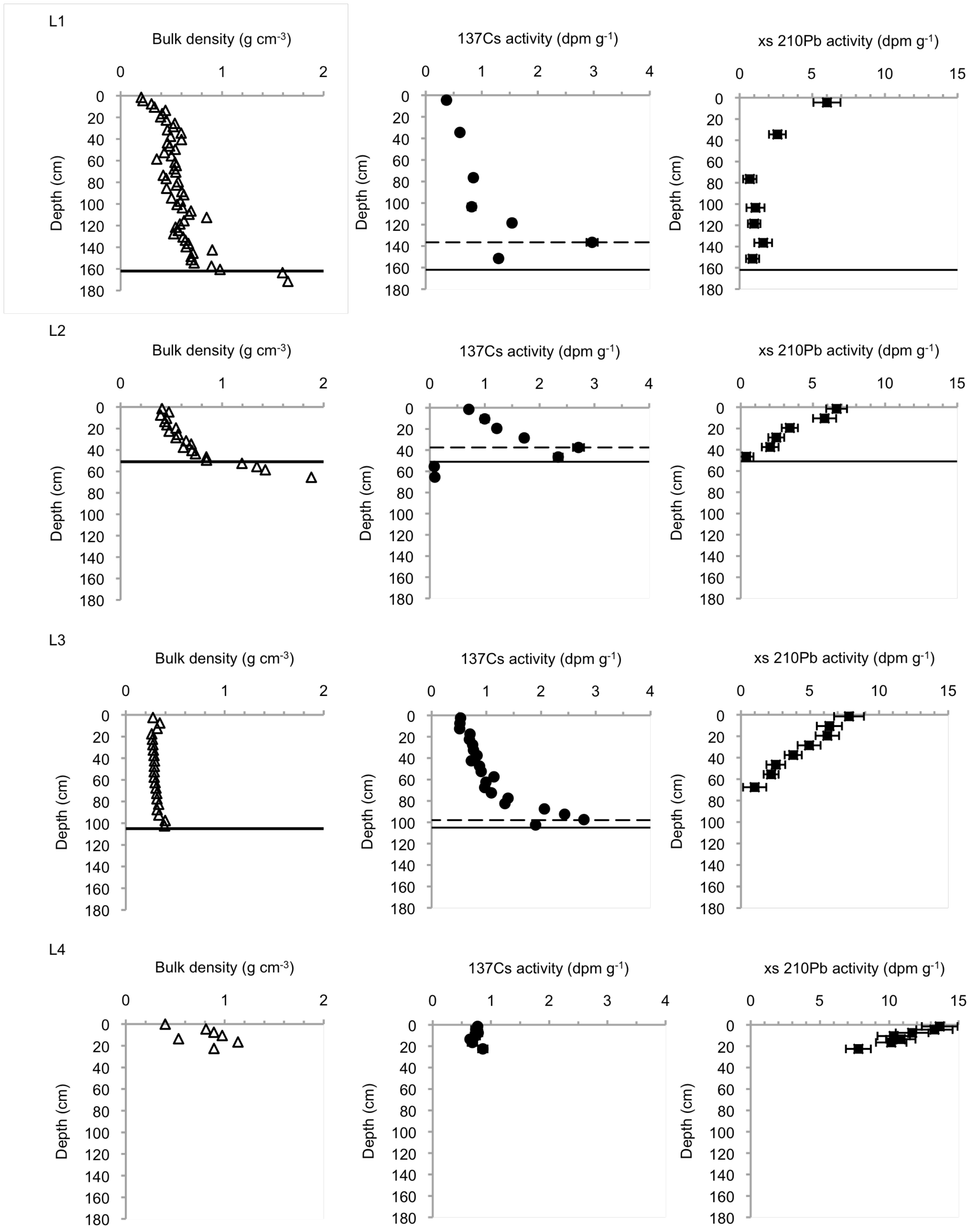

Figure 6. Sediment core profiles of bulk density and radioisotope activities from the eight reservoirs. Solid horizontal lines indicate the pre-impoundment surface (no line indicates the core did not capture the pre-impoundment surface). Dashed lines in ^{137}Cs graphs represent the depth of peak activity. The ^{210}Pb profile for L3 is from a second core collected at the same location.

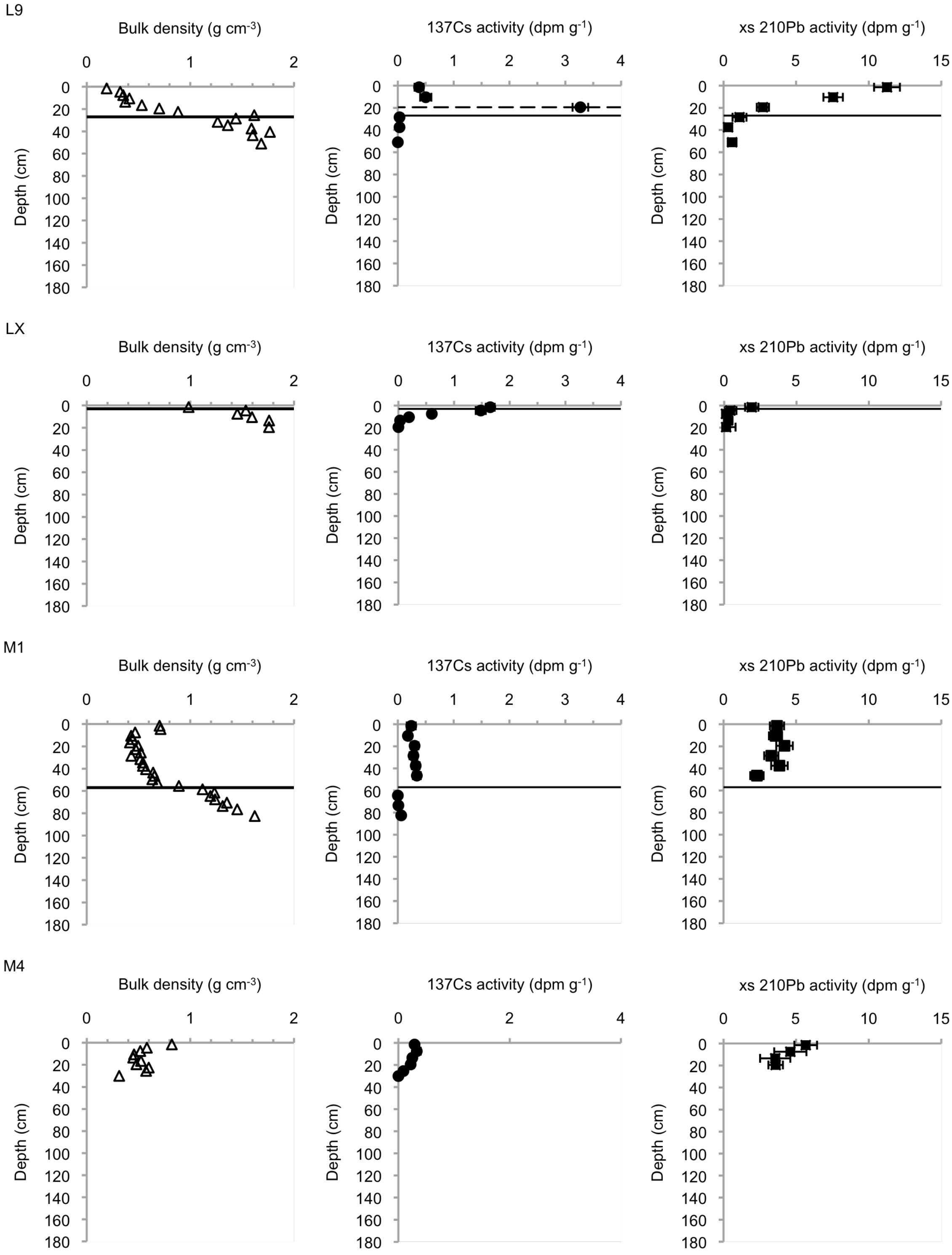

Figure 6 (continued). Sediment core profiles of bulk density and radioisotope activities from the eight reservoirs. Solid horizontal lines indicate the pre-impoundment surface (no line indicates the core did not capture the pre-impoundment surface). Dashed lines in ^{137}Cs graphs represent the depth of peak activity.

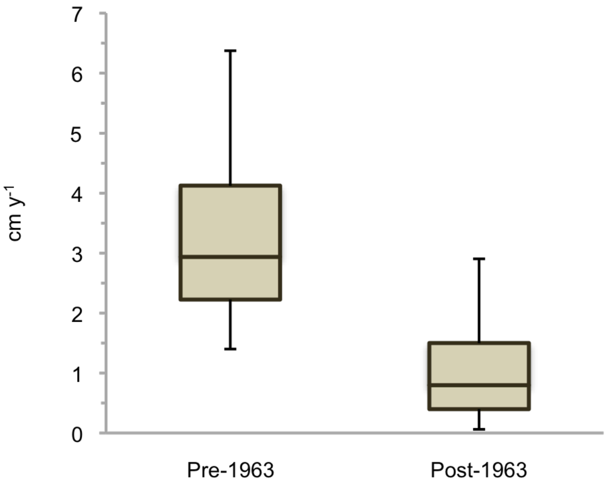

Figure 7. Linear sedimentation rates derived from ^{137}Cs activities. Summary comparison of pre-1963 and post-1963 rates.

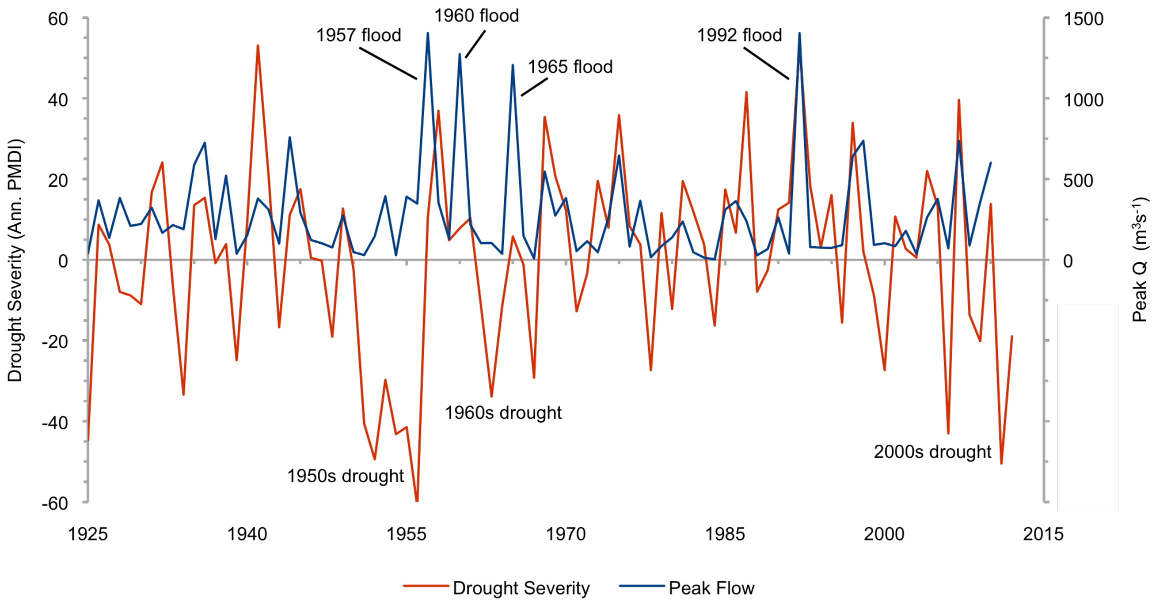

Figure 8. Chronology of regional drought (annual Palmer Modified Drought Index) and peak flows on the Lampasas River.