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Banded vegetation-dune development during the Medieval Warm Period and 20th century, Chihuahuan Desert, New Mexico, USA

Stacey L. Weems[1] and H. Curtis Monger[2,†]

[1]*U.S. Forest Service, Salt Lake City, Utah 84138 USA*
[2]*Department of Plant and Environmental Sciences, New Mexico State University, Las Cruces, New Mexico 88003-8003 USA*

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Abstract. With the advent of systematic high-resolution satellite photography, striking geometric shapes of banded vegetation several km^2 in size, but not apparent from the ground, have been documented for many areas of the arid and semiarid world. Banded vegetation, in which dense perennial vegetation alternates with bands of bare soil may originate from geomorphic processes, ecological self-organization, or human land use. In the Chihuahuan Desert of New Mexico prominent arc-shaped bands of vegetation and dunes occur along the contact of a piedmont slope (bajada) and basin floor. The origin and chronology of this banded vegetation-dune complex was investigated using early aerial photography (1936–1942), landscape photography (1918), vegetation and soil surveys (1858, 1918), soil stratigraphy, $^{13}C/^{12}C$ ratios, and ^{14}C dating. These methods reveal two periods of eolian deposition. The first began in the Medieval Warm Period (ca. AD 900–1300) and was followed by a period of landscape stability during the Little Ice Age (ca. AD 1500 to 1850). The second began in the late-1800s when widespread desertification occurred throughout the American Southwest. Banded vegetation was initiated after formation of erosional scarplets that functioned as obstacles upon which eolian sand accumulated, thus becoming a dam to overland flow and causing strips of vegetation to form. Banded vegetation in this study is an emergent pattern produced by a coupled ecologic-geomorphic-climatic system. The stratigraphic record produced by this system enables us to compare current ecological responses to climate change with baseline prehistorical responses to climate change.

Key words: carbon isotopes; climate change; desert geomorphology; emergence; Little Ice Age; Medieval Warm Period; object-based image analysis; positive feedback mechanisms; soil radiocarbon dating.

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† **E-mail:** cmonger@nmsu.edu

Introduction

Banded vegetation covers large areas in the arid and semiarid regions of Australia, southern Africa, the Sahel in east and west Africa, and the Americas (d'Herbès et al. 2001). These striking patterns of bands, arcs, or waves have been given names like tiger bush, mulga groves, and mogote (Tongway and Ludwig 1990, Montaña 1992, Valentin et al. 1999). The bands are typically oriented parallel to elevational contours on low gentle slopes where runoff occurs as sheet flow instead of incised drainage (Tongway and Ludwig 2001).

The origins of banded vegetation may result from the degradation of a continuous vegetative cover (Thiéry et al. 2001) or coalescence of smaller patches into more continuous linear zones (Bryan and Brun 1999). A critical aspect about the origin of banded vegetation is its starting points (Valentin et al. 2001). Did banded vegetation start forming in historical times as a result of human land use (Kéfi et al. 2007) or earlier in the Holocene as a result of climatic shifts (d'Herbès et al. 2001, Bowman et al. 2007)?

Once formed, the maintenance of banded vegetation can be generalized as follows. Bare zones have physical crusts with low infiltration rates that initiate runoff as thin overland sheetflow, where up to 80% of rainfall can be lost as runoff (Greene 1992). Obstructions to this flow cause sediment deposition, deep water infiltration, and deeply rooted vegetation (Ludwig et al. 2005). The runoff-obstruction-sedimentation-infiltration process is a positive feedback mechanism that maintains vegetated bands that receive not only runoff water from bare zones upslope but also nutrients from plant litter, animal dung, and seeds that have fallen onto the bare zones. Soils in vegetation bands in Australia, for example, had three times more nutrients than bare zones upslope (Tongway and Ludwig 1990).

As a scientific phenomenon, the formation of banded vegetation can be considered a complex adaptive system because it contains diverse agents that are connected, and interdependent, which causes the system to be adaptive (Holland 1975, Page 2006, Peters et al. 2006). It is also an example of "emergence" in which local-scale interactions and positive feedbacks generate striking large-scale patterns (Anderson 1972, Bestelmeyer et al. 2006, Rietkerk and van de Koppel 2007, Scanlon et al. 2007, Nicholas et al. 2009).

Banded vegetation reported in this study occurs as prominent bands along the contact of the lower piedmont slope and basin floor (Figs. 1 and 2). Bands are arc-shaped with a linear dune (1 to 2 m high) that follows the curvature of erosional scarplets. Parallel to and in contact with the dune on the upslope side is a robust zone of grass, primarily tobosa grass (*Hilaria mutica*), which thins upslope until it contacts its associated bare zone.

The two objectives of this study were to investigate (1) the chronology of the banded vegetation-dune complexes and (2) propose a mechanism for their development. Chronology was investigated by testing two hypotheses: that the banded vegetation formed during the late-1800s/early-1900s when widespread desertification began in the Southwestern U.S.A. (Bryan 1925, Buffington and Herbel 1965, Waters and Haynes 2001) versus the hypothesis that banded vegetation formed prehistorically during a period of high erosion in the Holocene (Antevs 1955, Irwin-Williams and Haynes 1970, Gile and Hawley 1966, Gile 1975, Hawley 1993, Buck and Monger 1999). Mechanism of formation was investigated by testing two hypotheses: that banded vegetation is inherited from underlying soil properties, such as abrupt texture differences (d'Herbès et al. 2001) or buried dune microtopography preserved in the subsoil (Zonneveld 1999) versus the hypothesis that erosional scarplets in the study area function as obstacles upon which wind-blown sand accumulates (Bagnold 1971, Greeley and Iversen 1985, Lancaster 1995), which, consequently, becomes a barrier to overland flow.

Methods

The average annual rainfall between 1915 and 1995 at the Jornada Experimental Range headquarters, three kilometers northwest of the study site, was 245 mm with a standard deviation of 85 mm. This includes sporadic monsoonal summer thunderstorms, winter frontal system, and long periods of drought, most notably the drought of the 1950s (Wainwright 2006). Most precipitation (52%) occurs July through September. Annual potential evaporation is 2204 mm. Average annual temperature is 14.7°C with a standard deviation of 0.6°C. The dominant wind direction is from the west-southwest and is greatest in March, April, and May, with the highest average wind speed occurring in April (12.4 km/hr, 7.8 mi/hr) (Wainwright 2006). No helicoidal winds have been reported for this area (e.g., Gillette and Chen 2001). Occasional gusts can reach 145 km/hr (90 mi/hr).

The physiography of the study site is Basin and Range with mountains oriented north-south and separated by structural basins filled with Cenozoic sediments (Hawley 2005) produced by the Rio Grande Rift tectonic system that has been active since the Oligocene (Seager et al. 1984).

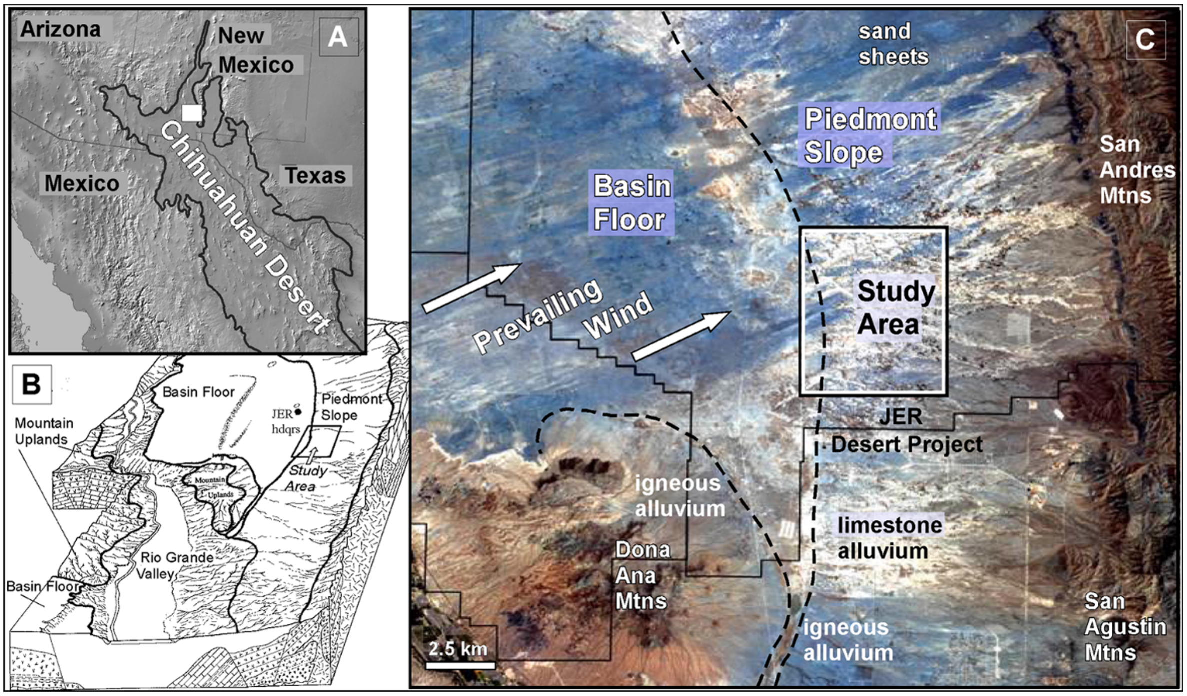

Fig. 1. (A) White square locates study conducted at the Jornada Experimental Range. (B) Illustration of the geomorphic setting and location of study area at the contact of the piedmont slope and basin floor. (C) Landsat image of study area vicinity and parent materials.

The study site is located along the contact of the basin floor and piedmont slope descending from the San Andres Mountains (Fig. 1). Slopes range from 2% in the eastern part of this contact to 1% in the western part (Gile et al. 1981). The alluvial parent material of the piedmont slope is derived from Paleozoic sedimentary rock, mainly limestone, which grades to the basin floor composed of fluvial sediments deposited by the ancestral Rio Grande during the early Pleistocene (Mack et al. 1996). Basin floor soils have strongly-developed profiles, most of which have stage V petrocalcic horizons. Horizons above the petrocalcic (A, Bt, Btk) have in many areas been truncated by wind erosion and reworked as eolian deposits. The mineralogy of the eolian sediments in the study site is predominantly quartz with minor amounts of feldspars and lithic fragments that can be traced laterally to their source on the basin floor (Monger 2006).

The dominant vegetation at the study site is tobosa grass (*Hilaria mutica*) which makes up the dense bands of vegetation. Tarbush (*Flourensia cernua*) and burrograss (*Scleropogon brevifolius*) are sparsely distributed across the lower portions of the bare runoff zones. The dunes are mainly occupied by grass (*Sporobolus* spp.), yucca (*Yucca elata*), and tarbush sparsely distributed within largely barren areas of sand.

To investigate chronology, evidence that the banded vegetation-dune complex formed historically was obtained from aerial photography taken in 1936 and 1942. The 1942 photographs were used to compare shrubs, grass, and bare ground in 1942 with 2003 imagery using ERDAS Imagine 9.0, georectified 1996 digital orthoquad (DOQ) images, and eCognition 4.0 (Laliberte et al. 2004, Definiens 2006, Weems 2007). Additional evidence was based on 1918 landscape photography obtained from the Rio Grande Historical Collection at New Mexico State University, a 1918 soil survey by the U.S. Forest Service in collaboration with the U.S. Bureau of Soils (Veatch 1918), and 1858 vegetation maps from Buffington and Herbel (1965) and Gibbens et al. (2005) created from land survey notes that included comments on the condition of the land (Gile 1966).

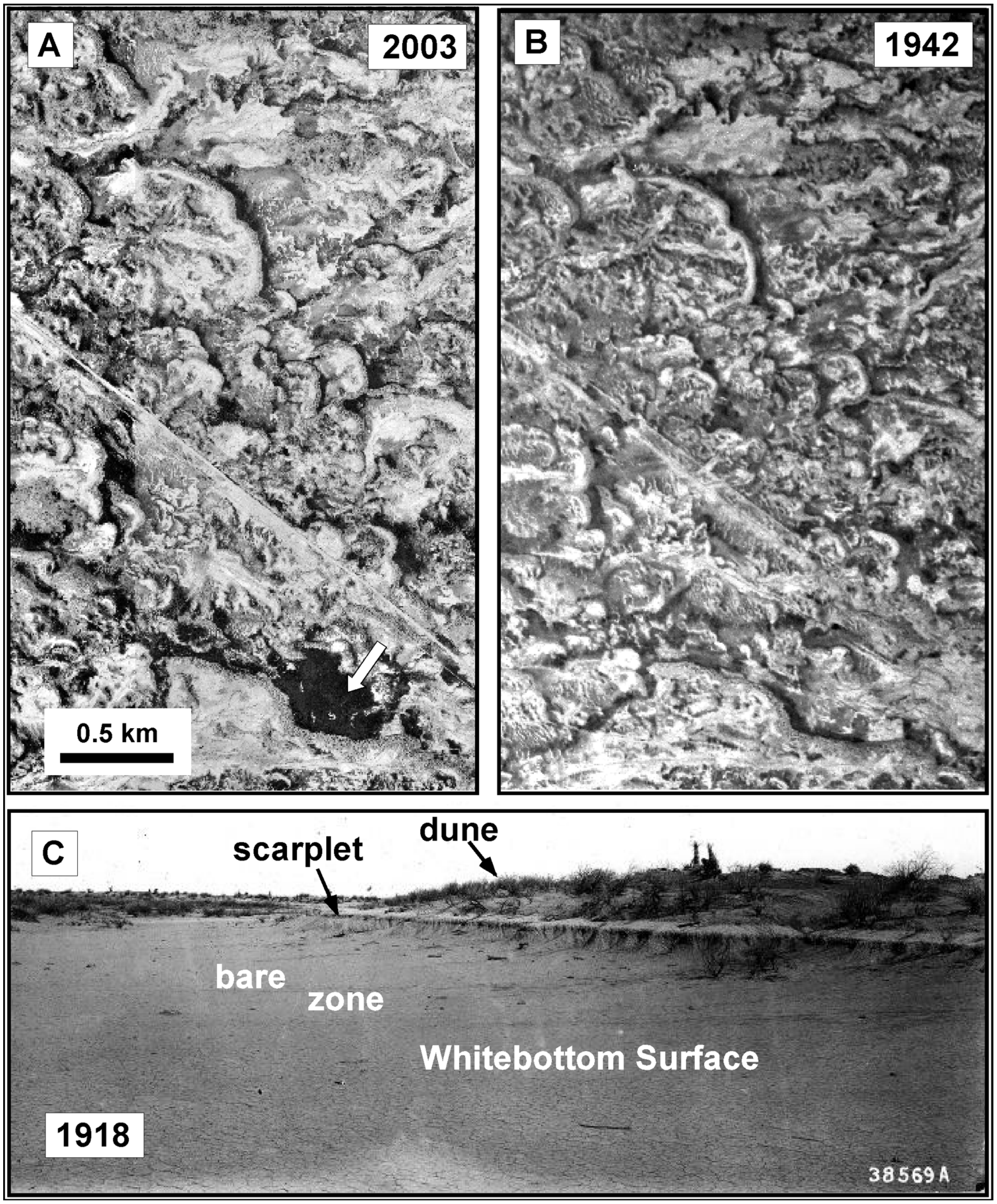

Fig. 2. (A) Aerial photograph of banded vegetation in the study area in 2003. Dark strips are dense zone of tobosa grass. Bare ground = 24.0%; shrubs = 60.7%; grass = 15.3% (Weems 2007). (B) Aerial photograph of banded vegetation in 1942. Bare ground = 8.7%; shrubs = 54.3%; grass = 37.0%. (C) 1918 landscape photograph of bare zone, scarplet, and linear dune that runs parallel to scarplet. Banded vegetation zone is behind dune. Photograph courtesy of Rio Grande Historical Collections, New Mexico State University. Note white arrow in 2003 photograph pointing to increased size of dense tobosa grass as compared to 1942.

Evidence that the banded vegetation-dune complex formed prehistorically was based on soil stratigraphy and geomorphic surfaces (Daniels et al. 1971) as revealed by a transect of five backhoe trenches, $^{13}C/^{12}C$ ratios, and ^{14}C dating. Names of geomorphic surfaces are from the neighboring Desert Project (Fig. 1C) and include the Organ III, II, and I surfaces, the Jornada II surface, the Whitebottom surface, and sand dune surfaces of Historical age (Ruhe 1964, Gile 1966, Ruhe 1967, Gile and Hawley 1968, Hawley 1975, Blair et al. 1990, Gile 1999, Monger et al. 2009). Although Spanish explorers passed through the region in the late 1500s, "Historical" as used geomorphically in this region has been assigned a date beginning in AD 1850 (Gile et al. 1981).

In contrast to soils deposited prehistorically, which are characterized by pedogenic development, soils deposited historically have stratified, fresh-appearing C-horizon material in both the topsoil and subsoil and the absence of macroscopic carbonate filaments or coatings (Gile 1966, Blair et al. 1990). The youthful age of these deposits is based on land survey notes and buried artifacts (Buffington and Herbel 1965, Gile 1966, Monger 1995). In addition, ^{14}C dating at the study site was used to date a buried charcoal hearth, a buried carbonized leaf, the wood of a dead tarbush shrub, soil organic matter, and pedogenic carbonate. For soil organic matter samples, a bulk soil sample was collected from the A horizon in each trench 20 cm below the modern or buried land surface.

$^{13}C/^{12}C$ ratios were also measured to provide information about the abundance of C_4 versus C_3 plants that accompanied dune deposition. Pedogenic carbonate analyzed for $\delta^{13}C$ was collected from all five profiles at 20 cm intervals from the surface to approximately 200 cm and sent to the stable isotope laboratory at New Mexico Tech University (Socorro, NM) and analyzed with a Finnigan MAT Delta E mass spectrometer (Ringoes, NJ) using Oz-tech gas standards. Carbon isotope ratios are presented in δ-notation:

$$\delta^{13}C(‰) = \frac{R_{SAMPLE} - R_{STD}}{R_{STD}} \times 10^3 \qquad (1)$$

where R is the $^{13}C/^{12}C$ ratio of the sample. Results are reported relative to Vienna Pee Dee Belemite *Belemnitella americana* (Craig 1957, V-PDB) by calibration with the NBS-19 standard. The fraction of carbon in the soil organic matter derived from C_4 plants (F_{C4}) was estimated by the equation (Boutton et al. 1999):

$$F_{C4} = \frac{\delta_{SAMPLE} - \delta_{C3}}{\delta_{C4} - \delta_{C3}} \qquad (2)$$

where δ_{SAMPLE} is the $\delta^{13}C$ of the bulk soil sample, δ_{C3} is the average $\delta^{13}C$ value of the C_3 components, δ_{C4} is the average $\delta^{13}C$ value of the C_4 components. End-member values of −25‰ were used for C_3 plants and −14‰ for C_4 plants (Boutton et al. 1999). The C_4 fraction based on pedogenic carbonate was estimated by the equation:

$$F_{C4} = \frac{\delta_{SAMPLEpedcarb} - \delta_{C3pedcarb}}{\delta_{C4pedcarb} - \delta_{C3pedcarb}} \qquad (3)$$

where $\delta_{SAMPLEpedcarb}$ is the $\delta^{13}C$ of the carbonate in the soil sample, $\delta_{C3pedcarb}$ is the average $\delta^{13}C$ value derived from the C_3 components, $\delta_{C4pedcarb}$ is the average $\delta^{13}C$ value derived from the C_4 components. End-member values of −12‰ were used for C_3 plants and +2‰ for C_4 plants (Quade et al. 1995).

Horizon nomenclature of the National Cooperative Soil Survey (Soil Survey Staff 2010) was used to describe soil profiles with the exception of a lower case "v" which was used to signify a vesicular A horizon in the bare-zone profile. The "K" horizon was used to identify the stage III calcic horizon (Gile et al. 1966). Complete soil characterization data of the dune and underlying buried soil can be found at http://soils.usda.gov for pedon number S09NM013011.

Results and Discussion

Aerial photography provides evidence that banded vegetation existed in the study area by the early to mid 1900s, although the amount of bare ground was less than today (Fig. 2). Landscape photography in 1918 also shows a banded vegetation-dune-scarplet landscape very similar to today's landscape (Fig. 2C). The soil survey (also in 1918) describes the study area as follows (Veatch 1918):

> ". . . in many places the wind has worked in front of the scarplets, since they face west, sweeping out the sediment until quite bare, and at the same time has deposited sand on the edge of the higher terrace. . . ."

Vegetation maps constructed from 1858 survey notes describe the area as being 15–55% dominated by tarbush (Buffington and Herbel 1965). A later reconstruction from the 1858 notes describes the area as consisting of tarbush, fourwing saltbush, and Yucca (Gibbens et al. 2005).

Backhoe trenching along the 100-meter transect revealed laterally continuous A, Btk, and K horizons of the Jornada II soil that could be traced beneath the dune until they were truncated by the Whitebottom surface (Fig. 3F). Taxonomically, the soil associated with the Jornada II surface is the Headquarters series, a fine-loamy, mixed, thermic Ustic Calciargid (Herbel and Gile 1973, Herbel et al. 1994, Gile et al. 2003, Soil Survey Staff 2010). The Whitebottom surface is an eroded variant of the Headquarters series. The dune is classified as the Pintura series: sandy over loamy, mixed, thermic Typic Torripsamment.

Based on crosscutting and superposition principles, the Jornada II surface is the oldest surface at the site because it is buried by the sand dune and is crosscut by the Whitebottom surface. The age of the Jornada II surface and its soil—which are the same since both date from the approximate time that sedimentation stopped and soil development started—is estimated to be late to middle Pleistocene, probably 75,000 to 150,000 years BP based on the amount of carbonate accumulation as constrained by volcanic ash and paleomagnetic dates (Mack et al. 1993, 1996, Gile 2002).

The age of the dune ranges from present (because deposition is still occurring) to the time when sand started accumulating on the Jornada II surface, which was after 760 BP based on the age of the charcoal hearth beneath the dune (Fig. 3F, Table 1). At 20 cm above the Jornada II surface a faint hiatus is present as indicated by a very weakly-developed A horizon 5 cm in thickness. The ^{14}C age of the soil organic matter in the weakly-developed A horizon is 780 BP. This date is not an absolute age, but rather a weighted mean of all the various organic compounds. A carbonized leaf in the very weakly-developed A horizon yielded a radiocarbon date of 650 BP (Fig. 3C, Table 1). The deposit of sand above the leaf is therefore younger than that age. The stratum of eolian sand between the Jornada II surface and leaf lacks the fresh-appearing sedimentary structure that is diagnostic of Historical deposits, but instead has weakly-developed peds characteristic of the Organ III allostratigraphic unit. In marked contrast, the eolian sand above the leaf has distinct sedimentary structure diagnostic of the Historical deposits found in coppice dunes and sandsheets in the region.

The age of the Whitebottom surface ranges from present (because it is still forming by erosion) to at least 310 BP (the age of the tarbush wood) since the Whitebottom surface must have been in existence for the tarbush to grow on it (Fig. 3F). The date of 310 BP is not considered an absolute age, but rather a general age representing an average of all carbon within the stem.

Radiocarbon ages of soil organic matter (SOM) in the A horizon are progressively older when traced from the tobosa grass (Fig. 3E) westward beneath the dune (Fig. 3F). Likewise, $\delta^{13}C$ values become more depleted when traced laterally beneath the dune. Based on $\delta^{13}C$ values of SOM and Eq. 2, the percentage of C_4 vegetation drops from roughly 100% in the tobosa soil at Profile E (Fig. 3E) to 67% beneath the dune at Profile B (Fig. 3B). Estimated percentages of C_4 based on pedogenic carbonate and Eq. 3 show fairly close agreement to SOM estimates except for the tobosa grass (Profile E) where the estimated percentage of C_4 vegetation was 69% based on pedogenic carbonate rather than 100% based on SOM (Table 2). The ^{14}C date at Profile A (Fig. 3A) is older than the other samples, which accords with previous studies that have shown samples deeper in soil profiles (in this case a truncated Btk horizon) typically have older ages because they receive less modern carbon (Wang et al. 1996).

The $\delta^{13}C$ values of the Btk and K horizons were also traced laterally. Both horizons show a dominance of C_4 vegetation (Table 2). The Btk has an overall average isotopic value of $-0.9‰$ (± 0.32, $n = 20$) and an estimated percentage of C_4 of 79% $\pm$ 2.3. The $\delta^{13}C$ values of the K horizon have an overall average isotopic value of $-0.3‰$ (± 0.18, $n = 12$) with an estimated percentage of C_4 of 83% $\pm$ 1.3 (Table 2).

Concerning the mechanism of formation, a few kilometers to the south of the study site in the same landscape position the soils are still the Headquarters series with scarplets, but without

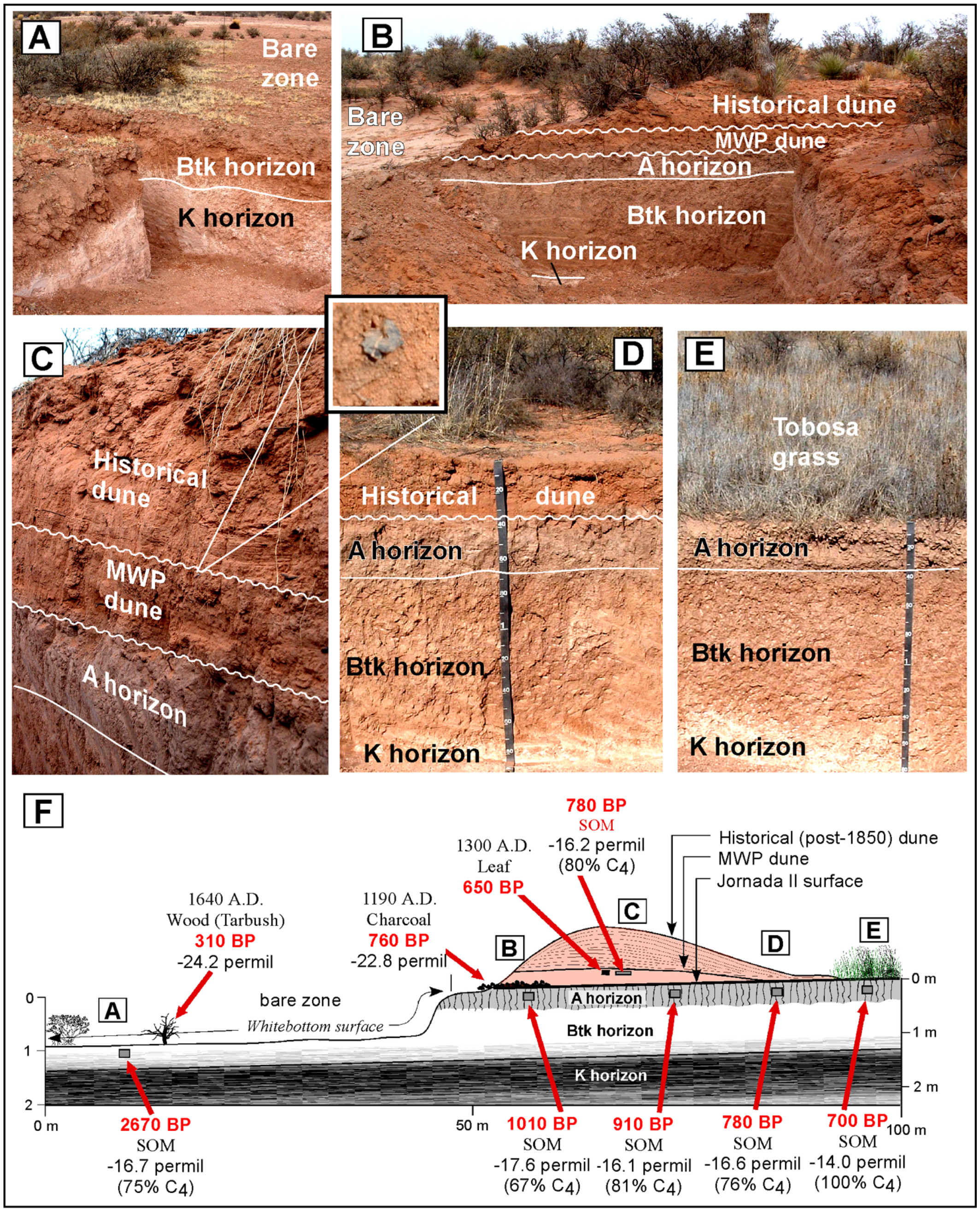

Fig. 3. Photographs of sample collection sites from soil profiles (A–E) and their location along a 100-meter transect (F). Cross-section diagram also illustrates soil stratigraphy, geomorphic surfaces, radiocarbon dates, and carbon isotope values. Width of ^{14}C-dated leaf is 1 cm.

dunes (Gile et al. 2003). Scarplets in this sand-deficit area range from about 20 cm to rarely more than 2 m and have miniature erosional footslopes grading to smooth grass-covered surfaces that receive increments of Whitebottom sediments from upslope (Gile et al. 1981). Still farther south a few more kilometers along the piedmont slope the parent material changes from

Table 1. Radiocarbon ages and $\delta^{13}C$ for shrub, leaf, charcoal, and soil organic matter (SOM) and pedogenic carbonate (Carb) for Profiles (Pr-) A to E.

Sample	Lab no.	Measured radiocarbon age	$\delta^{13}C$ (‰)	Conventional radiocarbon age	Calibrated age† (1 SD)	Calibrated‡ average age
Tarbush wood	Beta-215375	290 ± 40 BP	−24.2	300 ± 40 BP	1520–1650 AD (430–300 BP)	1640 AD 310 BP
Leaf	Beta-215368	N/A	N/A	670 ± 130 BP	1250–1410 AD (700–540 BP)	1300 AD 650 BP
Charcoal	Beta-215367	830 ± 40 BP	−22.8	860 ± 40 BP	1160–1230 AD (790–720 BP)	1190 AD 760 BP
SOM-Pr-E	Beta-215370	520 ± 40 BP	−14.0	700 ± 40 BP	...§	...§
SOM-Pr-D	Beta-215371	640 ± 40 BP	−16.6	780 ± 40 BP	...	...
SOM-Pr-C	Beta-215372	760 ± 40 BP	−16.1	910 ± 40 BP	...	...
SOM-Dune	Beta-215369	640 ± 40 BP	−16.2	780 ± 40 BP	...	...
SOM-Pr-B	Beta-215373	890 ± 40 BP	−17.6	1010 ± 40 BP	...	...
SOM-Pr-A	Beta-215374	2530 ± 40 BP	−16.7	2670 ± 40 BP	...	...
Carb-Pr-B	Beta-215366	5860 ± 40 BP	−1.1	6250 ± 40 BP	...	...

† Calibrated radiocarbon ages to calendar years using the Pretoria Calibration Procedure (Talma and Vogel 1993) and INTCAL98 (Stuiver et al. 1998).
‡ Intercept between uncalibrated conventional radiocarbon age and calibrated curve time scale.
§ Not converted to calendar years because soil organic matter is an open system containing a conglomeration of organic substances, each of which could have its own dendrochronological calibration.

limestone alluvium derived from the San Andres Mountains to igneous alluvium derived from the San Agustin Mountains (Fig. 1C). In the igneous alluvium the scarplets cease to exist and erosion patterns contrast markedly with that seen in the silty material to the north. As noted by Ruhe (1967), "scarplets have a distinct relationship to limestone alluvium which has silty textures that have a tendency to seal, have platy structure, low infiltration, and greater runoff." In contrast, soils in igneous alluvium are loamy with higher infiltration rates and are not prone to scarplet formation.

North of the study area a few kilometers along the lower piedmont slope, major sand sheets have blown from the basin floor (Ritchie et al. 2003, Monger et al. 2006), but there are no scarplets (Fig. 1C). Although some silty outcrops occur in this region, most are buried by sand. In contrast to the banded vegetation patterns, vegetation in this region of abundant sand flux consists of uniformly distributed mesquite coppice dunes, a pattern similar to the coppice dunes in the basin floor.

Summary and Conclusions

Like banded vegetation in general, those at the Jornada Experimental Range form on slopes with low gradients, have bands oriented perpendicular to slope and wind direction, and have bare zones composed of silty-textured soils that form physical crusts with low infiltration rates that gives rise to overland flow (d'Herbès et al. 2001, Tongway and Ludwig 2001). Unlike typical banded vegetation, the ones in this study have an arc shape with a linear dune (1 to 2 m high) that follows the curvature of scarplets. This type of banded vegetation-dune complex is a subset of the more common types, being most similar to the "grassy microdune and band" of d'Herbès et al. (2001).

Timing of banded vegetation-dune development

We find no evidence to support the hypothesis that the banded vegetation began forming *exclusively* during the widespread desertification event of the late-1800s/early-1900s. Earliest aerial photography show banded vegetation in 1936, although the amount of bare ground has increased in 75 years (Fig. 2). In addition, landscape photography shows banded vegetation, scarplets, and dune in 1918, which are also described in the 1918 soil survey (Veatch 1918). Although vegetation maps of 1858 do not mention the banded vegetation per se, they describe tarbush, presence of Yucca, and absence of good grass (Buffington and Herbel 1965, Gibbens et al. 2005), which accords with banded vegetation today.

Instead, we find evidence to support the hypothesis that the banded vegetation-dune complex began forming prehistorically. Sand accumulation started after 760 BP (the age of

Table 2. Isotopic values of pedogenic carbonate (p carb) and soil organic matter (SOM), and estimated percent of C_4 vegetation occupying the land surface during soil formation.

Horizon depth (cm)	$\delta^{13}C$ p carb (‰)	C_4 p carb† (%)	$\delta^{13}C$ SOM (‰)	C_4 SOM‡ (%)
Profile A				
AvBtk				
0	−1.70	74		
Btk				
20	−0.98	79	−16.7	75
40	−0.72	81		
K				
60	−0.32	83		
80	−0.20	84		
100	−0.22	84		
120	−0.42	83		
140	−0.02	86		
160	−0.33	83		
Profile B				
Dune				
0	−1.64	74		
20	−0.71	81		
40	−2.22	70		
A				
60	−2.34	69	−17.6	67
80	−1.81	73		
Btk				
90	−1.10	78		
100	−1.24	77		
120	−1.16	77		
140	−0.88	79		
160	−0.86	80		
180	−0.99	79		
K				
200	−0.27	84		
220	−0.32	83		
Profile C				
Dune				
0	−2.22	70		
20	−0.59	82		
40	−0.65	81		
60	−1.42	76		
80	−1.69	74		
100	−1.60	74		
A				
120	−1.66	74	−16.1	81
Btk				
140	−1.34	76		
160	−1.16	77		
180	−0.73	80		
Profile D				
Dune				
0	−2.90	65		
20	−4.97	50		
A				
40	−2.04	71	−16.6	76
60	−2.48	68		
Btk				
80	−1.46	75		
100	−1.35	76		
120	−0.60	81		
140	−0.95	79		
160	−0.70	81		
K				
180	−0.20	84		
Profile E				
A				
0	−3.27	62		
20	−2.32	69	−14.0	100
40	−1.45	75		
Btk				
60	−1.26	77		
80	−0.86	80		
100	−0.47	82		
120	−0.35	83		
140	−0.38	83		
K				
160	−0.58	82		
180	−0.67	81		

Table 2. Continued.

† C_4 estimates using pedogenic carbonate were based on a $\delta^{13}C$ value of +2‰ for 100% C_4 vegetation and −12‰ for 100% C_3 vegetation (Eq. 3).

‡ C_4 estimates using SOM were based on a $\delta^{13}C$ value of −14‰ for 100% C_4 vegetation and −25‰ for 100% C_3 vegetation (Eq. 2).

the charcoal) but before 650 BP (the age of the leaf). The radiocarbon ages of SOM, useful only as a general indicator of soil age, do not contradict this interpretation. The age of the Whitebottom erosional surface, which continues to form today, is older than 310 BP (the age of the tarbush). Thus, sand accumulation followed by formation of banded vegetation probably started during the latter part of the Medieval Warm Period, perhaps in the extraordinarily dry period of the AD 1200s, long indentified by dendrochronology (Schulman 1956).

Still, the late-1800s/early-1900s desertification event and current erosion have left their mark. In fact, the Historical sand deposit is more than twice the thickness of the Medieval Warm Period deposit (Fig. 3C). As such, the relative magnitudes of these deposits may be a rough measure of terrestrial response to environmental change in the Chihuahuan Desert region.

Mechanism of formation

We find no evidence to support the hypothesis that the banded vegetation pattern is inherited from underlying soil properties. Backhoe trenching and profile description reveal the Headquarters Series which was truncated by the Whitebottom erosional surface on the downslope side and buried by sand on the tread of the upslope side which produced the Pintura Series

(Fig. 3).

Instead, we find evidence to support the hypothesis that erosional scarplets function as an obstacle upon which eolian sand accumulates as linear dunes parallel to arcuate scarplets. Scarplets are antecedent to dunes, not vice versa. Evidence for this can be seen by moving south a few kilometers to the zone without sand flux where arcuate scarplets exist, but without dunes (Ruhe 1967: plate 2, Gile et al. 1981: p. 183). In this region there are minor areas of banded vegetation, but they exist as the mogote type, which is more common and more developed farther south in Mexico (Montaña et al. 2001). A few kilometers north of the study area in a high sand flux zone, there are no scarplets, no arcuate dunes, and no banded vegetation, indicating that a combination of arcuate scarplets and dunes are prerequisite for the formation of this particular type of banded vegetation.

The ^{14}C dates of SOM is progressively older as the buried A horizon is traced beneath the dune westward (Fig. 3F). We attribute this to progressively less modern carbon being added to the A horizon as it became buried. The ages of SOM in the A horizon, suggest that the scarplet, dune, and banded vegetation are slowly migrating upslope, which is typical for banded vegetation based on a variety of evidence, including younger plants on the upslope edge of bands in contrast to old and dead plants on the downslope edge (Montaña et al. 2001, Tongway and Ludwig 2001). The velocity of this migration is on the order of 50 m per 300 years based on the age differences between the SOM at Profile E and B (Fig. 3F).

The $\delta^{13}C$ values of both SOM and pedogenic carbonate indicate that this landscape has been dominated by C_4 vegetation throughout the late Quaternary (the age of the Jornada II surface). Although there is one shrub in the Jornada region that uses C_4 photosynthesis (the fourwing saltbush—*Atriplex canescens*), we interpret the isotopic signal to represent C_4 grasses because the organic matter in the buried A horizons is similar to carbon at the tobosa site at Profile E (i.e., fibrous grass roots instead of larger-diameter woody roots). The approximate percentage of C_4 vegetation based on SOM $\delta^{13}C$ values drops from 100% in the tobosa profile to 68% beneath the dune, suggesting carbon inputs from C_3 Yucca and tarbush growing on the dune. The $\delta^{13}C$ values of pedogenic carbonate also indicate this landscape has been dominated by C_4 vegetation. A recent isotopic study in central New Mexico concluded that pedogenic carbonates have an isotopic bias toward C_4 vegetation (Breecker et al. 2009). However, pedogenic isotopic values at Profile E show a C_3, not C_4, bias because this site is currently a dense C_4 tobosa grassland, has a SOM isotopic signature indicating 100% C_4, yet pedogenic carbonate indicates only 69% C_4 (Table 2).

Based on soil stratigraphy, tracing the banded vegetation north and south into greater and lesser amounts of sand and scarplet occurrence, carbon isotopes, and ^{14}C dates, we propose the following model for banded vegetation formation in the study area.

1. A pre-eroded, pre-banded vegetation state consisted of uniform C_4 grassland, probably tobosa grass, which covered the Jornada II surface (Fig. 4A).
2. Around AD 1190 a charcoal hearth was deposited on the Jornada II surface (Fig. 4B). During this period, overland flow and erosion increased as bare ground increased. Because of the silty soil texture, vertical scarplets approximately 50–60 cm high began to form in an arcuate pattern on the lower piedmont slope. Sediments were carried from the scarplets downslope across the Whitebottom surface to the next runon zone containing dense grass.
3. After the formation of scarplets, sand began accumulating on the Jornada II surface owing to scarplets functioning as obstacles that creates turbulence to wind-blown sand (Fig. 4C). The origin of the sand, based on mineralogical tracing, is truncated soil horizons above petrocalcic horizons to the west on the basin floor. The linear dunes running parallel to the scarplets were now functioning as dams behind which overland flow and nutrients accumulated and dense bands of grasslands formed. The radiocarbon-dated leaf was deposited near the top of a MWP dune around AD 1300.
4. A period of greater vegetative cover, landscape stability, and soil formation occurred during the Little Ice Age (ca. AD 1500 to

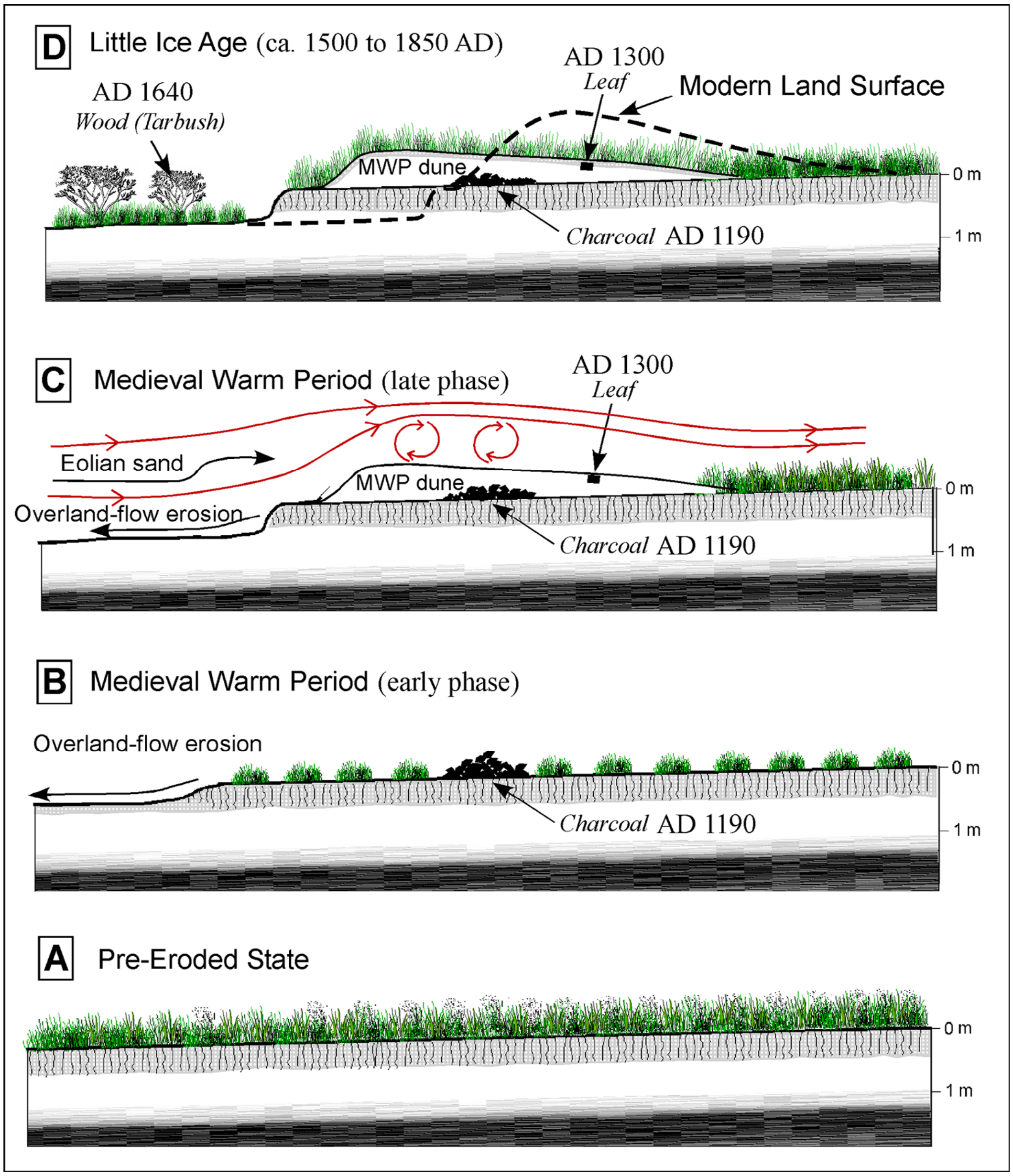

Fig. 4. Illustration showing proposed development of the scarplet-dune system. (A) Cross-section illustration of pre-eroded state. (B) Initiation of erosion and deposition of charcoal during Medieval Warm Period. (C) Sand accumulation following scarplet formation and deposition of radiocarbon-dated leaf. (D) Landscape stability during Little Ice Age and expansion of tarbush and grass across Whitebottom surface.

1850) when denser stands of grass and tarbush (based on the AD 1640 ^{14}C date) expanded across the Whitebottom erosional surface (Fig. 4D).

5. With the advent of the late-1800s/early-1900s desertification event, vegetative cover declined throughout the region. In the neighboring basin floor and throughout

the piedmont slope both wind and water erosion increased. Sand deposition resumed and buried the MWP dune, a process that continues today (Fig. 4D).

Implications

The desertification event of the late-19th and 20th centuries corresponded to the introduction of large numbers of domestic livestock (Campbell 1929, Fredrickson et al. 1998, Havstad et al. 2006). During this period large areas of the American Southwest converted from grasslands to woody shrublands in part because of selective herbivory, seed dispersal, and trampling of vegetation (Gardner 1951, Buffington and Herbel 1965, Schlesinger et al. 1990). The loss of grass cover gave rise to widespread erosion and formation of coppice dunes (Gile 1966, Herbel et al. 1972, Gibbens et al. 1983) and increased spatial and temporal heterogeneity of soil moisture and nutrients (Wright 1982, Schlesinger et al. 1996, Peters et al. 2010). This historical desertification event was preceded by the climatic effects of the Little Ice Age, an interval (ca. AD 1500 to 1850) when expansion and subsequent fluctuation of alpine glaciers occurred throughout the world's high mountain systems (Lamb 1977, Porter 1986, Thompson et al. 1986) and cooler and moister conditions occurred in the American Southwest (Fritts 1976, Neilson 1986, Castiglia and Fawcett 2006). The Little Ice Age, in turn, was preceded by the Medieval Warm Period, an interval (AD 900 to 1300) of elevated temperatures in northern Europe (Lamb 1977), but also recognized in the western and midwestern United States (Schulman 1956, LaMarche 1974, Stine 1994, Broecker 2001, Cook et al. 2004, Meyer and Frechette 2010).

The vegetation patterns at the Jornada are a legacy of these three climatic periods which reinforces the assertion that an understanding of prehistorical vegetation is important for understanding current patterns of vegetation. Banded vegetation is an emergent feature produced by linked ecologic-geomorphic-climatic processes. It may therefore preserve a record of how ecosystems responded to climate changes in the past. The fact that the Historical dune in this study is twice as thick as the MWP dune may be a useful baseline against which current rates of desertification can be compared.

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