PRELIMINARY ASSESSMENT OF CLIMATIC CHANGE DURING

LATE WISCONSIN TIME, SOUTHERN GREAT BASIN AND

VICINITY, ARIZONA, CALIFORNIA, AND NEVADA

By W. Geoffrey Spaulding, Stephen W. Robinson, and Frederick L. Paillet

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[Plates in Pocket]

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CONVERSION FACTORS

Multiply SI units	By	To obtain inch-pound units
millimeter (mm)	0.03937	inch
meter (m)	3.281	foot
kilometer (km)	0.6214	mile (mi)
gram (g)	0.03527	ounce, avoirdupois
kilogram	2.2046	pounds
degree Celsius (°C)	$F = 9/5°C + 32$	degree Fahrenheit

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ABSTRACT

Concentration and relative abundance of plant macrofossils illustrate compositional variations in samples from the Eleana Range-2 packrat (*Neotoma* sp.) midden. Nine macrofossil assemblages spanning 6,500 radiocarbon years record local vegetational changes in the southern Great Basin of Nevada during the last one-half of the late Wisconsin glacial age. The vegetation of the Eleana Range-2 site, on a south-facing slope at 1,810 meters altitude, was characterized by limber pine (*Pinus flexilis*) and steppe shrubs, from before 17,100 radiocarbon years before present to shortly after 13,200 radiocarbon years before present. Changes toward a more xerophytic plant association at the site began by 16,000 radiocarbon years before present, culminating in a major change to pinyon-juniper (*Pinus monophylla-Juniperus osteosperma*) woodland between 13,200 and 11,700 radiocarbon years before present. This latest Wisconsin woodland differed from that of the present primarily in the composition of its understory.

The climatic reconstruction for the late full glacial episode (17,000 to 15,000 radiocarbon years before present) that is proposed to account for limber pine-shrub vegetation in the Eleana Range is characterized by increased winter precipitation, perhaps 70 percent greater than the present winter precipitation, and very little summer rainfall. Average annual precipitation probably did not exceed present annual precipitation by more than 40 percent, a climate somewhat drier than most pluvial climatic reconstructions. Annual temperatures during the full glacial were perhaps 6° to 7° Celsius lower than the present average, and evidence exists for winter temperatures much lower than those of the present. A major warming trend occurred between about 16,000 and 12,000 radiocarbon years before present and was largely concordant with major dessication of closed lakes in the southern Great Basin. By the latest Wisconsin (12,000 to 10,000 radiocarbon years before present), annual temperatures were within 2° Celsius of temperatures of the present.

A period of wetter conditions in the southern Great Basin during the latest Wisconsin may have incorporated increased precipitation during both the summer and winter, and lower temperatures during the winter, relative to the present. Resurgence of closed lakes in the southern Great Basin between 12,000 and 10,000 radiocarbon years before present and the persistence of mesophytic vegetation where desert now exists are explicable phenomena in accordance with a proposed latest Wisconsin pluvial event. The climate of this pluvial event was much different from that of the preceding full glacial episode. Its general nature is consistent with what would be expected from current models of astronomically induced climatic changes.

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INTRODUCTION

The southern Great Basin is a region of bioclimatic transition between subtropical deserts to the south and continental steppe deserts to the north. In the Sonoran Desert, south of about lat 34° N. (pl. 1), winters are mild, summers are hot, and the precipitation regime is bimodal, with distinct winter and summer rainy seasons. In the floristic Great Basin Desert, north of about lat 37° N. (approximately congruent with the northern two-thirds of the physiographic Great Basin) (Fenneman, 1931; Shreve, 1942), winters are cold, summers are shorter and cooler, and much of the annual precipitation falls during autumn and winter. The southern Great Basin lies between these two areas, encompassing most of the floristic Mojave Desert (pl. 1) (Rowlands and others, 1982). Much of the southern Great Basin is characterized by cool winters and hot, relatively dry summers, with about 75 percent of average annual precipitation occurring in fall through spring (Shreve, 1942; Beatley, 1975, 1976; Rowlands and others, 1982).

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Vegetational zonation is well-developed in the southern physiographic Great Basin. Abrupt environmental changes result in discrete plant communities and well-defined ecotones, particularly from low valleys to adjacent high mountains. Such abrupt environmental changes make ideal conditions for the study of vegetational response to changing climates. This research has included the analysis of present distributions of plant communities and local climates (Billings, 1949; Beatley, 1974, 1975; Rowlands and others, 1982) and paleoecological investigations (Mehringer, 1965, 1967; Wells and Berger, 1967; Spaulding, 1983a). Paleoecological studies also provide information on the nature of past climates and the chronology of climatic change during the late Quaternary. Such data can be compared to geohydrologic records of paleoclimates in the area. In particular, reconstructions of the timing and magnitude of variations in pluvial-lake levels provide sensitive chronologies of climatic change (Smith, 1979; Smith and Street-Perrott, 1983). The availability of several independent paleoecological and geohydrologic studies in the southern Great Basin provides an opportunity to refine current models of late Quaternary climatic changes in the American Southwest.

The southern Great Basin has been the focus for recent paleoenvironmental research using plant macrofossil assemblages from ancient packrat (*Neotoma* sp.) middens. More than 50 fossil sites from the Sheep Range and the vicinity of the Nevada Test Site in Clark and Nye Counties, Nevada, and adjacent Inyo County, California, provide records of the last 45,000 years of vegetational and climatic change (Spaulding, 1981, 1983a). Particularly informative data were obtained from sites yielding macrofossil chronosequences spanning parts of Wisconsin and Holocene time. One such record, from the Eleana Range in Nye County, Nevada, provides perspective on the nature of climates and of the chronology of the transition from the Wisconsin maximum to the beginning of the Holocene.

Geographic setting

The Eleana Range is a small mountain range of volcanic rocks of Tertiary age in south-central Nevada, in the southeastern Great Basin. Altitudes range from 1,500 m at the foot of the range to a maximum of 2,230 m. Many low mesas occur in the range, formed by beds of welded tuff capping less resistant units of nonwelded tuff. Rock shelters are common in exposures of nonwelded tuffs. The Eleana Range is adjacent to the floristic boundary of the Mojave and Great Basin Deserts (pl. 1). Regional vegetation is discussed by Beatley (1976), and botanical nomenclature follows her work in most instances. Shrub vegetation on the lower flanks of the range gives way to singleleaf pinyon pine-Utah juniper (*Pinus monophylla-Juniperus osteosperma*) woodland at higher altitudes. The lower altitudinal limit of woodland near the Eleana Range-2 site is between 1,750 and 1,850 m, depending on site orientation and exposure. In the vicinity of the packrat-midden site, at 1,810 m altitude, sparse pinyon-juniper woodland occupies the mesa, whereas sagebrush (*Artemisia tridentata, A. nova*) dominates in the alluvium-filled valleys (fig. 1). Common associates of pinyon pine and juniper include jointfir (*Ephedra viridis*), dwarf-goldenbush (*Haplopappus nanus*), and sagebrush. Between sparsely wooded mesas and the sagebrush-covered valleys is a third habitat characterized by shrub vegetation. Plants on talus slopes flanking the mesas include jointfir, Mormon-tea (*Ephedra nevadensis*), saltbush (*Atriplex canescens*), rabbitbrushes (*Chrysothamnus nauseosus, C. viscidiflorus*), and grasses (*Stipa comata, Sitanion hystrix, Bromus tectorum*). Mesic habitats in the vicinity support isolated stands of Gambel oak (*Quercus gambelii*) and thickets of skunk-bush (*Rhus trilobata*).
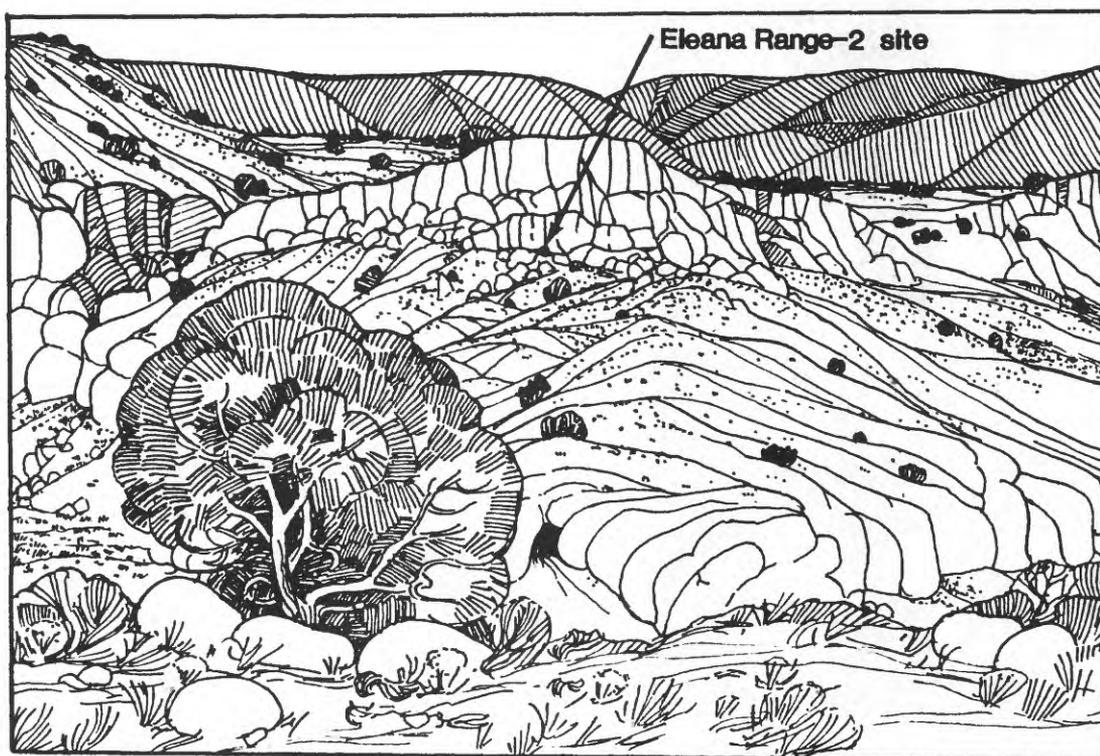
Figure 1.--Physiography and vegetation of the Eleana Range-2 site. Widely spaced pinyon pines and Utah juniper occupy the mesa top and occur as isolated individuals on the shoulders of the mesa.

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Acknowledgments

METHODS

The Eleana Range-2 packrat midden is a stratified accumulation of plant debris and packrat feces, about 1.2 m high and 2 m wide (fig. 2).
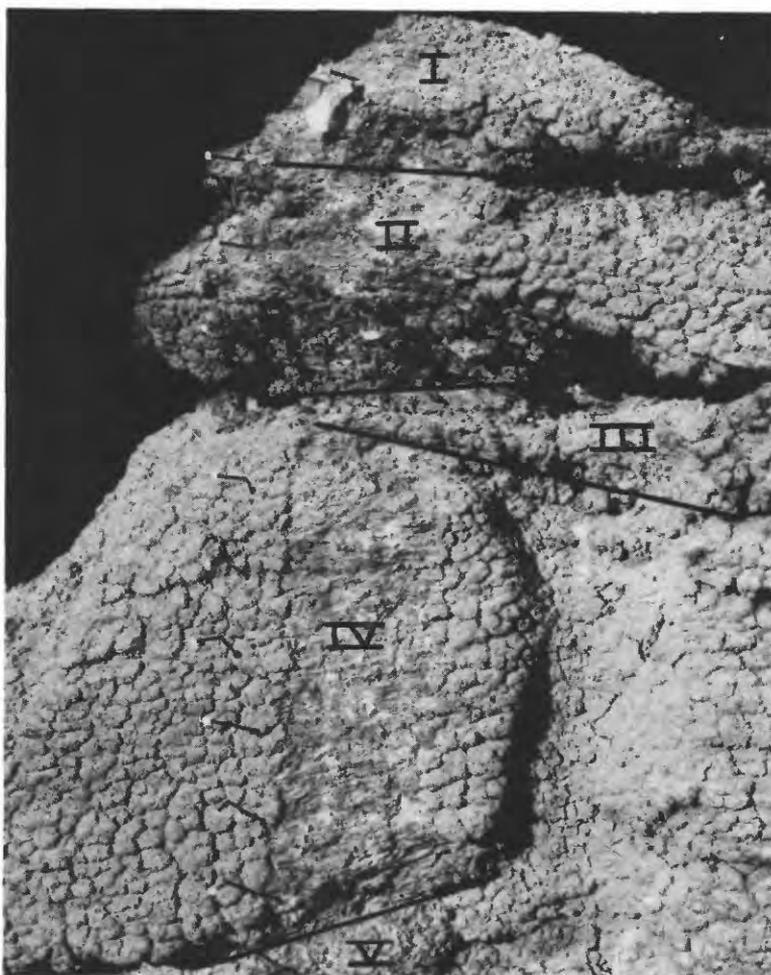

Figure 2.--West one-half of the Eleana Range-2 packrat midden, showing major stratigraphic units (I to V). The weathering rind has been removed from the sample column prior to removal of the samples. Nails to the left of the sample column generally are at 100-millimeter intervals.

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The distance from the midden's exterior face to its contact with the back wall of the rock shelter is about 1.3 m. The deposit is thoroughly impregnated with crystallized packrat urine (amberat) that has cemented the midden into a well-indurated mass. As with most ancient packrat middens, a convoluted weathering rind composed of dust, degraded midden debris, and amberat armors the exterior of the midden (fig. 2). Prior to sampling, the weathering rind was chipped from a 300-mm wide vertical section, and the stratigraphy was sketched. Samples for macrofossil analysis were collected along a single vertical column, cut 100 to 200 mm into the deposit, from the exterior face. Excavation proceeded by removing discrete blocks from the midden with a hammer and cold-chisel. The position of these samples relative to the stratigraphy of the deposit was carefully noted.

In the laboratory, the exterior of each midden sample was trimmed, cleaned, and sketched if it was to be subsampled. After weighing, the indurated samples were immersed in warm water for 3 to 5 days to dissolve the cementing amberat and release the macrofossils. The disaggregated samples were washed through nested soil sieves, dried, and then weighed again. Plant macrofossils were sorted from the mass of organic debris, identified, and counted or weighed (Spaulding, 1983a).

Analytic methods are not standardized in packrat-midden studies, a relatively new technique in Quaternary paleoenvironmental research (Wells, 1976; Spaulding and others, 1983). Graphic displays of time-series data and attempts to enumerate macrofossil abundance recently have been published (Betancourt and Van Devender, 1981; Cole, 1983; Spaulding, 1983a, b). Two enumerative methods are used in this study: (1) Tree macrofossils are reported in grams per kilogram of sample; and (2) shrub and succulent macrofossils are reported as numbers of identified fragments per sample.

In packrat-midden analysis, many inferences of past vegetational conditions depend on the assumption of proportional representation. Common or abundant macrofossil types in a packrat-midden assemblage are presumed to represent those plant species that were common or abundant in the local paleovegetation. Comparison of modern vegetation with plant debris in currently active or recently indurated packrat middens indicate that, in most instances, this presumption is true (Van Devender, 1973; Wells, 1976; Cole, 1981, 1983; Spaulding, 1981, 1983a). Studies of modern packrat middens indicate that a packrat normally is an indiscriminate collector, gathering those plant species that are most readily available near the midden site. Correlations between the abundance of plant fragments in a midden sample and the importance of plant species near the site are not perfect, but they are sufficient to allow reconstruction of vegetation to the community-type level. Forest, woodland, and desert-scrub vegetation can be discriminated on the basis of macrofossil assemblages, and because most of the fossil taxa can be identified to species, plant-association types can be determined. For example, limber-pine woodland can be differentiated from ponderosa-pine (*Pinus ponderosa*) forest, and juniper woodland can be distinguished from pinyon-juniper woodland (Van Devender, 1973; Wells, 1976; Spaulding and others, 1983). When multiple midden samples from a discrete area and from a limited time frame yield abundant remains of the same plant species, those species are assumed to have been dominant in the local paleovegetation.

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The area covered by a foraging packrat is limited to the immediate vicinity of the midden site (Raun, 1966; Stones and Hayward, 1968; Cranford, 1977). Long-distance transport of plant fragments to a midden is considered extremely unlikely. Estimates of effective foraging distance range from 30 to 100 m (Van Devender, 1973; Phillips, 1977; Cole, 1981; Spaulding, 1981). Fragments of plant species that occur as much as 100 m away occasionally occur in modern middens, but most foraging activity probably takes place within 30 m of the midden site. Cole (1981, 1983) and Spaulding (1981, 1983a) compared the plant fragments in recent middens with plants growing within 30 m of the midden site, and they concluded that the percent similarity (as measured by Sorensen's index, I; Sorensen, 1948) normally is between 65 and 90 percent. Wells (1976) observed that, as a consequence of predator pressure, fitness in packrats should be maximized within the minimum home range sufficient to support an individual packrat.

Macrofossil abundance in packrat-midden samples traditionally has been reported in terms of subjectively assigned values (Wells and Berger, 1967; Van Devender, 1973; Mead and Phillips, 1981; Spaulding, 1981). While the method is more time-efficient, it yields imprecise estimates of macrofossil abundance, and it is a technique less likely to provide comparable data when used by different analysts. Recognizing this, recent studies use either weights or counts of macrofossils to report plant-fragment frequency (Cole, 1981, 1983; Spaulding, 1983a, b; Betancourt and Davis, 1984). When packrat middens accumulate in woodland vegetation, the needles of pine or the twigs of juniper commonly number into the many thousands per kilogram of midden. In such instances, weighing the macrofossils of arboreal species is more efficient than counting. A ratio of the weight of an arboreal taxon to the sample weight provides an estimate of macrofossil concentration. The weight is that of the washed sample, minus the weight of rocks, fecal pellets, and identified plant remains. Thus, these concentration values represent macrofossil abundance relative to a background of undifferentiated midden debris.

Numbers of identified specimens are used to report frequencies of shrub and succulent macrofossils in this study. The mass of such macrofossil types in the Eleana Range-2 samples usually was small, and it became practical to count macrofossils when those of each taxon generally numbered less than one thousand per sample. Relative changes in the number of shrub and cactus macrofossils are illustrated as percent variations of the total number of identified specimens counted in each sample (NISP), excluding arboreal taxa (NISP-AT).

Radiocarbon dating is essential for chronologic control in packrat-midden studies. Rates of midden accumulation are variable and episodic (Van Devender, 1973; King, 1976; Cole, 1981; Spaulding, 1981, 1983a). Normal superposition of midden strata cannot be assumed, and cases of reversed stratigraphy have been documented (Van Devender, 1973; Cole, 1981; Spaulding, 1981). Reversed superpositioning probably results from the accumulation of a younger midden under a ledge formed by an ancient, indurated deposit (Van Devender, 1973; King, 1976). Amberat saturation leads to the formation of another indurated mass beneath the first. Because rates of deposition vary, even samples from middens displaying regularly bedded, horizontal strata need to be radiocarbon dated to determine accurate age relationships.

Mummified organic debris from ancient packrat middens provides ideal material for radiocarbon dating. Twigs, foliage, seeds, and wood of several plant species, as well as packrat fecal pellets and finely macerated plant debris, were analyzed for carbon-14 (^{14}C) and carbon-13 (^{13}C) (table 1). Dating was performed on the cellulose component of these materials, extracted from cleaned plant debris by standard techniques (Sheppard, 1975, p. 11). Pretreatment of samples for radiocarbon analysis included multiple washes with hot distilled water and diluted potassium hydroxide, which remove any soluble contaminants from the plant debris. Uriniferous residue was considered unsuitable for radiocarbon assay, because amberat is mobile within middens and is a potential contaminant (Spaulding, 1983a).

Table 1.--*Samples and radiocarbon dates from the Eleana Range-2 packrat midden*

[Artr, *Artemisia* sec. Tridentatae; Cele, *Cercocarpus ledifolius*; fp, *Neotoma* fecal pellets; Juos, *Juniperus osteosperma*; ndl, pine needles; Pifl, *Pinus flexilis*; sd, seeds; tg, twigs; upd, unidentified plant debris; wd, wood]

Unit	Sample number	Radiocarbon date (±1σ)	Laboratory number[1]	Material dated	Weighted average[2]
I	ER-2(1)	10,620±120	USGS-876	Juos tg & sd	---
	ER-2(2)	11,700±85	USGS-914	Artr tg	---
II	ER-2(3)	13,180±65	USGS-884	fp	---
	ER-2(4)	14,090±320	A-2927	Pifl ndl	---
	ER-2(4)	15,300±130	USGS-875	Pifl wd	---
III	ER-2(5)	14,260±110	USGS-874	Pifl wd	---
IV	ER-2(6)	14,500±110	USGS-885	Pifl ndl	---
	ER-2(6)	14,830±80	USGS-886	fp	14,720±70
	ER-2(7)	14,760±100	USGS-887	fp	---
	ER-2(7)/ 2(8)	15,380±100	USGS-888	Cele wd	---
	ER-2(9)	15,890±120	USGS-889	fp	---
	ER-2(10)	17,100±130	USGS-890	fp	---
	ER-2(10)	17,100±70	USGS-1069	Pifl ndl	17,100±70
V	ER-2(11)	10,800±90	USGS-892	upd	---
	ER-2(11)	14,310±110	USGS-893	fp	---

[1]USGS--Radiocarbon Laboratory, U.S. Geological Survey, Menlo Park, California; A--Laboratory of Isotope Geochemistry, Department of Geosciences, University of Arizona, Tucson.

[2]Long and Rippeteau (1974).

In the Eleana Range-2 packrat midden, stratigraphic discontinuities are accompanied by bedded weathering rinds indicating hiatuses in deposition. Based on these conditions, four major units were defined (fig. 3). Radio-carbon dates ranging from 10,620±120 yr B.P. [radiocarbon years before the present (1950)] at the top of Unit I to 17,100±70 yr B.P. near the bottom of Unit IV (fig. 3; table 1) indicate generally normal superposition. Average deposition rate of the Eleana Range-2 midden is about 10 mm per 65 years. Deposition of Unit IV was faster, about 10 mm per 55 years, and deposition of Units I and II was slower, about 10 mm per 125 years. The lowest unit in the midden, Unit V [represented by sample ER-2(11)] was subjected to bioturbation. Less indurated than the units above, it contains a large number of insect exoskeletons, as well as mouse and pika (*Ochotona* sp.) fecal pellets. Radio-carbon dates from ER-2(11) are anomalously young and not concordant (table 1). The macrofossil assemblage from this stratum is a mixture of plant remains of varying ages, probably the result of animals tunneling under the deposit; this unit is not considered in this analysis. Abundant pika pellets in this sample indicate that pika were living under the Eleana Range-2 midden during the last part of the late Wisconsin. The nearest living population of pika occurs today in the Toquima Range of central Nevada, 190 km north-northwest of the Eleana Range (pl. 1) (Grayson, 1982; Thompson and Mead, 1982).

A different case of mixing is indicated by the date of 15,300±130 yr B.P. on a limber-pine branch from sample ER-2(4) (table 1). The specimen's radio-carbon age is more than 1,000 years older than associated limber-pine needles in the ER-2(4) assemblage, dated at 14,090±320 yr B.P. (table 1). This latter date is close to that expected, considering the bracketing dates on samples above and below ER-2(4) (fig. 3; table 1). Packrats occasionally incorporate older items into accumulating middens, and the piece of limber-pine wood probably was gathered into the deposit more than 1,000 years after it grew on a local tree. Phillips (1977) and Cole (1981) discussed the inclusion of older items in younger middens. A necessary prerequisite to this phenomenon is a fairly large rock shelter or cave where organic remains may be preserved for centuries.

LATE WISCONSIN VEGETATION

Maximum glacial climatic conditions were reached at about 18,000 yr B.P. in the northern hemisphere (CLIMAP, 1976; Gates, 1976). The oldest macro-fossil assemblages from the Eleana Range-2 packrat midden, ER-2(10) and ER-2(9), provide evidence for vegetational conditions toward the end of the Wisconsin maximum. Radiocarbon-dated at about 17,100 yr B.P. and about 15,900 yr B.P. (table 1), these assemblages contain abundant limber pine, a conifer presently restricted to high-altitude subalpine woodlands in the Great Basin. Needles and buds of this tree are the dominant macrofossil types in all Eleana Range-2 assemblages, except those from Unit I (pl. 2; table 2). Limber pine is extinct in the Eleana Range, and it also is absent from adjacent highlands. The closest population today occurs in the central Groom Range, about 60 km to the northeast, at altitudes more than 475 m above the Eleana Range-2 site (pl. 1) (Beatley, 1976).

MIDDEN UNITS   SAMPLE COLUMN   RADIOCARBON DATE

UNIT I

ER-2(1) — 10,620 ± 120

ER-2(2) — 11,700 ± 85

ER-2(3) — 13,180 ± 65

UNIT II

ER-2(4) — 14,090 ± 320

UNIT III

ER-2(5) — 14,260 ± 110

ER-2(6) — 14,720 ± 70

ER-2(7) — 14,760 ± 100

15,380 ± 100

UNIT IV

ER-2(8)

ER-2(9) — 15,890 ± 120

ER-2(10) — 17,100 ± 70

10,800 ± 90

14,310 ± 110

UNIT V

ER-2(11)

Tuff of shelter floor

DEPTH, IN CENTIMETERS

0   10   20   30   40   CENTIMETERS
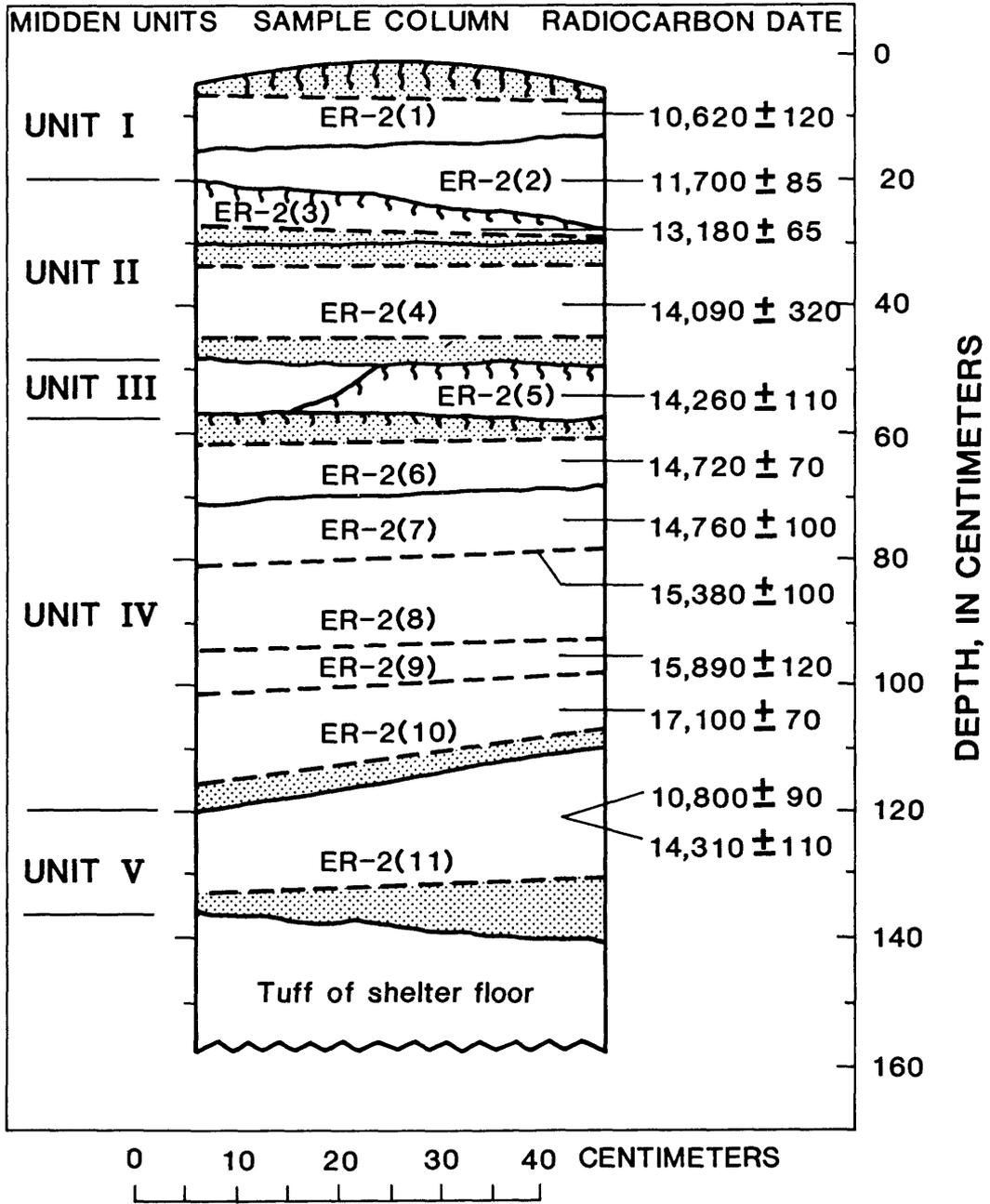
EXPLANATION

AREAS NOT SAMPLED   DISCONTINUITY WITH WEATHERING RIND

— — — LIMITS OF SAMPLE   ———— DISCONTINUITY

Figure 3.--Stratigraphic sketch of the Eleana Range-2 sample column with major midden units, samples, and radiocarbon dates (see table 1).

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Table 2.--*Plants from the vicinity of the Eleana Range-2 site and the Eleana Range-4 packrat middens*

[Relative abundance values for plant species within 30 meters of the site are in parentheses and indicate: 5, very abundant; 4, abundant; 3, common; 2, occasional; 1, rare. The modern Eleana Range-4 packrat midden is located 7 meters west of the Eleana Range-2 site. An X indicates a species present within 500 meters of the site; NISP-AT, number of identified specimens minus arboreal taxa; N, number of plant taxa; N_{ts}, number of species of trees, shrubs, and succulents; I, Sorensen's index of similarity (using trees, shrubs, and succulents only)]

Species	Vegetation (relative abundance)	ER-4	ER-2(1)	ER-2(2)	ER-2(3)	ER-2(4)	ER-2(5)	ER-2(6)	ER-2(7)	ER-2(9)	ER-2(10)
Approximate radiocarbon age, years	---	modern	10,600	11,700	13,200	14,100	14,300	14,600	14,800	15,900	17,100
						Macrofossil assemblages sample number and count (weight in grams)						
Agropyron sp.	(1)	---	---	---	---	3	---	---	---	---	---
Arceuthobium sp.	---	---	---	---	---	19	---	---	---	---	---
Artemisia nova	(1)	---	---	---	---	---	---	---	---	---	---
Artemisia tridentata	(X)	---	---	---	---	---	---	---	---	---	---
Artemisia sec. Tridentatae	---	---	30	45	153	178	317	100	139	111	172
Asteraceae undet.	---	---	82	21	146	335	95	41	71	43	25
Atriplex canescens	(4)	337	6	---	---	---	---	---	---	---	---
Boraginaceae undet.	---	---	24	41	29	4	23	9	13	8	8
Brickellia californica	(1)	---	---	---	---	---	---	---	---	---	---
Brickellia sp.	---	---	2	2	---	---	---	---	---	---	---
Bromus tectorum	(4)	38	---	---	---	---	---	---	---	---	---
Castilleja sp.	---	---	5	---	---	2	1	---	---	---	---
Cercocarpus ledifolius	---	---	114	---	390	106	25	1	6	139	10
Chamaebatiaria millefolium	---	---	4	11	14	56	90	61	54	20	24
Chrysothamnus nauseosus	(2)	8	15	19	44	48	102	24	29	55	35
C. viscidiflorus	(1)	8	9	2	82	184	111	43	109	24	20

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Table 2.--Plants from the vicinity of the Eleana Range-2 site and the Eleana Range-4 packrat middens--Continued

	Vegetation (relative abundance)	Macrofossil assemblages sample number and count (weight in grams)									
		ER-4	ER-2(1)	ER-2(2)	ER-2(3)	ER-2(4)	ER-2(5)	ER-2(6)	ER-2(7)	ER-2(9)	ER-2(10)
Cirsium sp.	---	---	1	---	---	---	---	---	---	---	---
Cowania mexicana	(X)	11	---	---	---	---	---	---	---	---	---
Cryptantha confertiflora	(2)	---	---	---	---	---	---	---	---	---	---
Cryptantha sp.	---	---	---	2	---	1	---	---	---	---	---
Ephedra nevadensis	(1)	---	---	---	---	---	---	---	---	---	---
E. viridis	(2)	31	---	---	---	---	---	---	---	---	---
Ephedra sp.	---	132	---	---	---	---	---	---	---	---	---
Eriogonum caespitosum	(X)	---	---	---	---	23	---	---	---	---	---
E. deflexum	(1)	---	---	---	---	---	---	---	---	---	---
E. microthecum	(2)	---	---	---	---	---	---	---	---	---	---
Eriogonum sp.	---	107	3	5	---	---	5	1	---	1	---
Euphorbia sp.	---	---	1	3	1	---	---	---	---	---	---
Festuca sp.	---	---	---	---	---	---	---	---	---	---	---
Forsellesia nevadensis	(X)	---	---	---	---	---	---	---	1	---	---
Gilia sp.	---	---	---	---	---	1	---	---	1	---	---
Gutierrezia microcephala	(X)	---	---	---	---	---	---	---	---	---	---
Haplopappus nanus	(X)	2	231	131	108	104	87	67	53	19	---
Heterotheca vilosa	(X)	---	---	---	---	---	---	---	---	---	---
Hilaria jamesii	(X)	---	---	---	---	---	---	---	---	---	---
Holodiscus microphyllus	---	---	4	---	7	20	5	2	11	2	1

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Table 2.--Plants from the vicinity of the Eleana Range-2 site and the Eleana Range-4 packrat middens--Continued

	Vegetation (relative abundance)	Macrofossil assemblages sample number and count (weight in grams)									
		ER-4	ER-2(1)	ER-2(2)	ER-2(3)	ER-2(4)	ER-2(5)	ER-2(6)	ER-2(7)	ER-2(9)	ER-2(10)
Juniperus osteosperma	(1)	8.3	8.8	1.3	<0.1	---	---	---	---	---	---
Lepidium sp.	---	---	---	---	1	---	---	---	---	---	---
cf. Leptodactylon pungens	---	---	32	1	---	2	5	2	2	15	3
Lesquerella sp.	---	---	9	---	1	---	1	3	---	1	1
Linum lewisii	(X)	3	---	---	---	---	---	---	---	---	---
Lupinus cf. argenteus	---	---	7	6	12	86	256	170	191	145	212
Machaeranthera canescens	(2)	---	---	---	---	---	---	---	---	---	---
Mirabilis froebelii	(1)	---	---	---	---	---	---	---	---	---	---
Machaeranthera sp.	---	---	1	---	---	---	1	---	---	---	---
Opuntia cf. erinacea	(X)	15	586	576	---	---	---	---	---	---	---
Opuntia sp.	---	---	---	---	---	1	---	---	---	---	---
Oryzopsis hymenoides	(2)	11	14	6	7	27	17	18	15	11	15
Penstemon sp.	(X)	---	1	---	2	---	---	3	21	2	2
Petradoria pumila	(X)	---	1	3	2	6	2	---	---	8	---
Phacelia crenulata	(1)	---	---	---	---	---	---	---	---	---	---
Phacelia mustelina	(2)	---	---	---	---	---	---	---	---	---	---
Phacelia sp.	---	29	---	---	---	---	---	---	---	---	---
Pinus flexilis	---	---	<0.1	<0.1	4.1	15.3	10.8	19.5	8.1	11.1	17.0
P. monophylla	(X)	0.5	0.5	3.4	---	---	---	---	---	---	---
Poaceae undet.	---	56	58	127	116	46	203	302	164	108	300
Prunus fasciculata	(X)	---	---	---	---	---	---	---	---	---	---
Purshia tridentata	---	---	65	40	---	24	12	18	4	1	---

Table 2.--Plants from the vicinity of the *Eleana Range-2* site and the *Eleana Range-4* packrat middens--Continued

	Vegetation (relative abundance)	Macrofossil assemblages sample number and count (weight in grams)									
		ER-4	ER-2(1)	ER-2(2)	ER-2(3)	ER-2(4)	ER-2(5)	ER-2(6)	ER-2(7)	ER-2(9)	ER-2(10)
Quercus gambelii	(X)	---	---	---	---	---	---	---	---	---	---
Rhus trilobata	(X)	2	---	---	---	---	---	---	---	---	---
Ribes cf. *velutinum*	---	---	2	---	---	2	3	---	---	---	---
Salvia columbarii	(X)	---	---	---	---	---	---	---	---	---	---
Salvia dorrii	---	---	99	88	---	---	---	---	---	---	---
Senecio multilobatus	(X)	---	---	5	---	---	---	---	---	---	---
Sitanion hystrix	(2)	4	---	---	3	---	---	---	---	---	---
Sphaeralcea cf. *ambigua*	(X)	4	1	1	---	---	---	---	---	---	---
Stanleya pinnata	(1)	---	---	---	---	---	---	---	---	---	---
Stipa comata	(3)	30	---	2	---	---	1	1	---	---	---
Symphoricarpos longiflorus	(1)	5	---	---	---	---	---	---	---	---	---
Symphoricarpos sp.	---	---	67	33	---	---	---	---	---	---	---
Tetradymia canescens	(3)	---	---	---	---	---	---	---	---	---	---
Tetradymia sp.	---	55	6	3	67	183	208	222	186	214	160
Yucca baccata	(X)	---	---	---	---	---	---	---	---	---	---
Weight, processed sample	---	19.3	23.9	27.8	27.0	38.6	12.2	14.4	61.8	42.0	50.5
NISP-AT	---	894	1,480	1,173	1,185	1,461	1,570	1,088	1,069	927	988
N	[1]23	23	33	27	20	25	24	20	17	20	16
N$_{ts}$	[1]11	14	20	16	11	14	12	11	11	11	9
I^2	---	[3]72	59	60	48	36	31	32	32	32	26

[1]Includes only those species occurring within 30 meters of the site.

[2]Comparison of macrofossil assemblages is with modern assemblage from the *Eleana Range-4* packrat midden.

[3]Comparison with trees, shrubs, and succulents growing within 30 meters of the site.

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Only a few woodland plants are associated with limber pine in the older assemblages from the Eleana Range-2 packrat midden. The rosaceous shrubs curl-leaf mountain mahogany (*Cercocarpus ledifolius*) and fernbush (*Chamaebatiaria millefolium*) presently occur in xerophytic woodland and chapparal in the central Great Basin (Linsdale and others, 1952; Brown and others, 1979). Arborescent mountain mahogany commonly occurs at the base of volcanic-rock outcrops in the northern Great Basin today, habitats similar to the Eleana Range-2 site at higher latitudes. Variable relative frequencies of mountain mahogany macrofossils (pl. 2) may reflect the occasional presence of single large shrubs in front of the Eleana Range-2 rock shelter.

Silvery lupine (*Lupinus* cf. *argenteus*) is a shrub that presently grows with sagebrush in sunny openings in the singleleaf pinyon pine-Utah juniper woodland of Pahute Mesa, within 10 km and 100 m altitude of the fossil site (pl. 1). Macrofossils of heliophytic shrubs such as silvery lupine, sage-brush, rabbitbrushes, and horsebrush (*Tetradymia* sp.) are important associates of limber pine in the Eleana Range-2 assemblages. Species of sagebrush, rabbitbrush, and horsebrush are common dominants of Great Basin steppe vegetation from valleys as low as 1,800 m to xeric mountain habitats higher than 3,000 m. Large percentages of these shrubs, coupled with the abundance of limber pine (pl. 2) indicate an open limber-pine steppe-shrub woodland at the Eleana Range-2 site (fig. 4) during the Wisconsin maximum, persisting until shortly after 13,200 yr B.P. General analogs to this vegetational type exist today in the central Great Basin, particularly at xeric sites above about 2,400 m altitude.

Other late Wisconsin packrat middens record subalpine conifer woodland in the southern Great Basin at altitudes generally above 1,800 m (fig. 5) (Mehringer and Ferguson, 1969; Thompson and Mead, 1982; Spaulding and others, 1983). Bristlecone pine commonly was a component of glacial-age subalpine vegetation on calcareous substrates. Midden sites from igneous substrates and localities outside the Great Basin provide no evidence for either Intermountain bristlecone pine (*Pinus longaeva*; Bailey, 1970) or Rocky Mountain bristlecone (*P. aristata* Englem.) (Thompson and Mead, 1982). The lower altitudinal limit of southern Great Basin subalpine conifer woodland varied, depending on substrate, aspect, mountain-mass size, and geographic position of the locality (Spaulding, 1981, 1983a; Wells, 1983). It also varied in response to short-term climatic fluctuations during the late Wisconsin (fig. 5) (Spaulding, 1981).

During the Wisconsin glaciation, subalpine conifer vegetation was the highest of two woodland zones in the southern Great Basin (fig. 5). Thermoph-ilous pygmy conifers, principally Utah juniper with some singleleaf pinyon pine, typified the lower woodland zone (Mehringer, 1965, 1967; Wells and Berger, 1967; Van Devender and Spaulding, 1979; Spaulding and others, 1983). Juniper and pinyon macrofossils are absent from all except the youngest Eleana Range-2 assemblages (pl. 2). This south-facing site, at 1,800 m altitude probably was well within the zone of subalpine conifer woodland during the Wisconsin maximum. In terms of general vegetational zonation during the full glacial period, the Eleana Range locality was above the local ecotone between thermophilous, pygmy-conifer woodland and subalpine woodland on volcanic substrate. In contrast, the juniper-subalpine conifer woodland transition was

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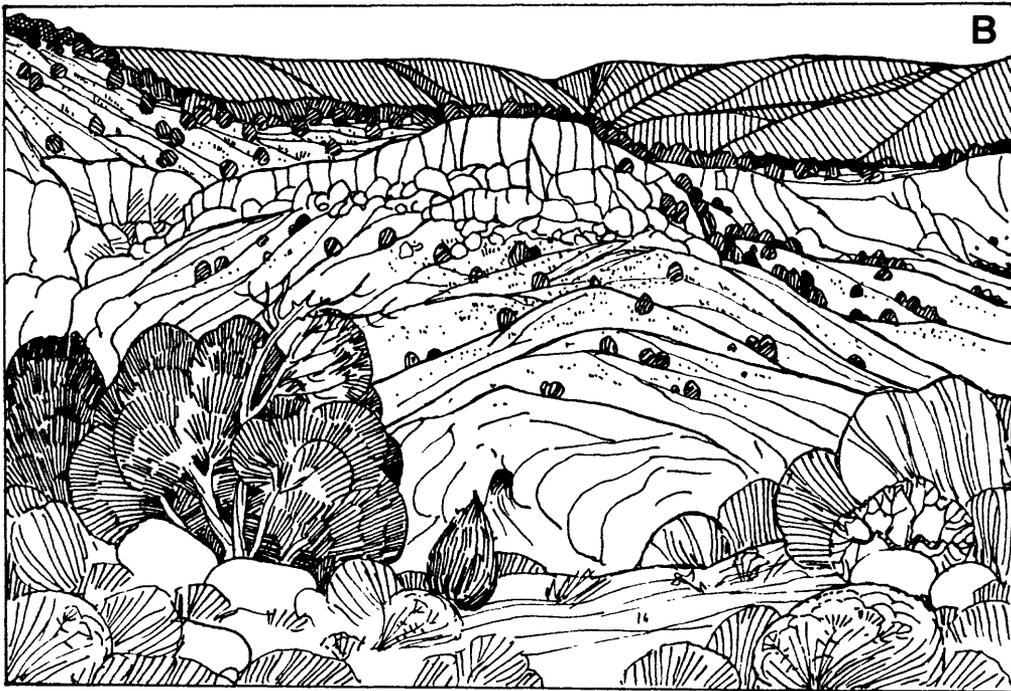
Figure 4.--Hypothesized vegetational conditions: A, during the
Wisconsin maximum (about 17,000 radiocarbon years before present),
and B, during the latest Wisconsin (about 11,000 radiocarbon
years before present) at the Eleana Range-2 site. Note that
while species composition differed radically between these two
communities (pl. 2), there was apparent continuity of the
physiognomy of the vegetation.
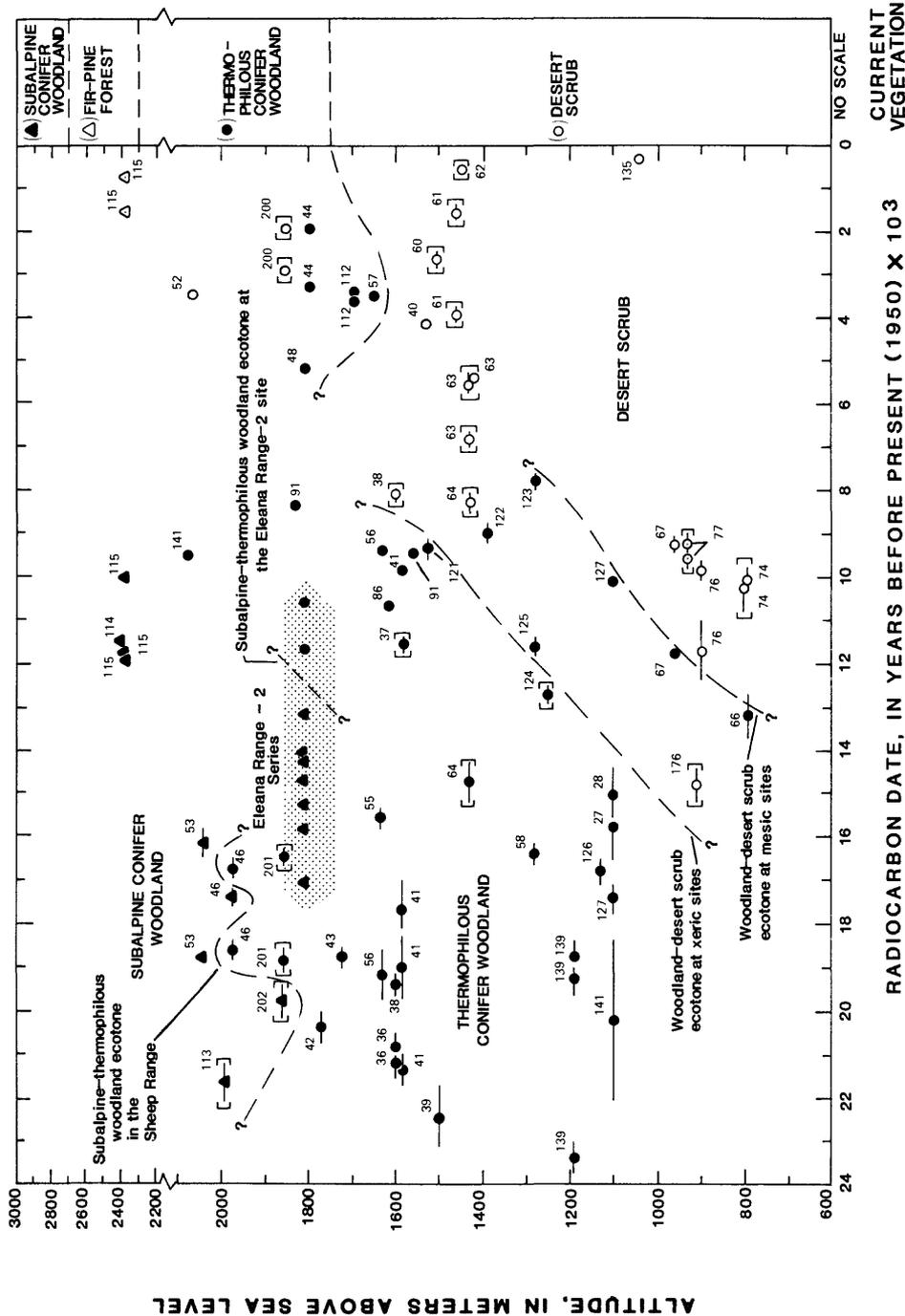
Figure 5.--Age, altitude, inferred paleovegetation from packrat-midden samples in the southern Great Basin north of latitude 36° N. Brackets indicate sites from xeric (southeast to west) aspects. All records are from calcareous substrate, except those from the Eleana Range-2 site, designated by the stippled zone. Position of ecotones (dashed lines) is approximate. See Supplementary Data section at the end of the report for key to site numbers, sample dates, and references.

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as high as 2,000 m altitude on calcareous substrates of the xeric Sheep Range, 100 km to the east-southeast (pl.1 and fig. 5) (Spaulding, 1981, 1983a). Soils derived from limestone and dolomite are effectively xeric, and vegetational zones usually occur at higher altitudes on mountains of carbonate rocks. Mesophytic plant species and woodland and forest vegetational zones occur at lower elevations on volcanic soils. Mountain ranges of carbonate rocks similar in size to the Eleana Range generally do not support woodland under modern climatic conditions (Wells and Berger, 1967; Beatley, 1976).

VEGETATION AT THE CLOSE OF THE WISCONSIN AGE

Between about 13,200 and 11,700 yr B.P., a pronounced alteration of vegetation occurred at the Eleana Range locality. Evidently, by 11,700 yr B.P., subalpine woodland gave way to thermophilous woodland typified by Utah juniper, pinyon pine, and prickly-pear cactus (*Opuntia* cf. *erinacea*). The two youngest samples from the Eleana Range-2 packrat midden reflect conditions after this change took place. The ER-2(1) and ER-2(2) macrofossil assemblages dated to 10,620±120 yr B.P. [ER-2(1)] and 11,700±85 yr B.P. [ER-2(2)] contain abundant macrofossils of pinyon pine, Utah juniper, and prickly pear (pl. 2). Limber pine and most of the shrub species that were common in the older samples are present in only trace quantities (table 2). The final decrease of sagebrush, horsebrush, and silvery lupine appears to have been largely coincident with the decrease in limber pine (pl. 2). Prickly pear and desert sage (*Salvia dorrii*) were understory plants that became established within the same time as pinyon pine and Utah juniper. Bitterbrush (*Purshia tridentata*) appears earlier, near the end of the Wisconsin maximum in sample ER-2(9), dated to 15,890±120 yr B.P. (pl. 2; table 1). Desert sage and bitterbrush do not occur near the site today; desert sage occurs primarily at lower altitudes, whereas bitterbrush generally is part of higher altitude woodland communities.

The change from limber pine-steppe shrub woodland to pinyon pine-Utah juniper woodland was profound. During the last millenia of the late Wisconsin, woodland zones were retreating upslope in the southern Great Basin (Mehringer, 1967; Spaulding, 1983a); the Eleana Range-2 packrat midden documents the passage of the subalpine-thermophilous woodland ecotone through the area between 13,200 and 11,700 yr B.P. (fig. 5). The local manifestation of upslope woodland retreat was the demise of subalpine and steppe species at this site, and the establishment of thermophilous trees and shrubs (pl. 2). At lower altitudes (below 1,300 m) during the last millenia of the late Wisconsin, pygmy-conifer woodland changed to desert vegetation (fig. 5) (Spaulding, 1983b).

Differences exist between the pinyon-juniper woodland of the latest Wisconsin and the woodland that currently exists at the Eleana Range locality (figs. 1 and 4). During the last 10,600 years marked changes occurred in the composition of the understory. Comparisons between the ER-2(1) and ER-2(2) macrofossil assemblages and the modern Eleana Range-4 packrat midden, collected nearby, reflect the Holocene expansion of saltbush and jointfir or Mormon-tea (pl. 2). Prickly-pear cactus decreased to a point where it is presently rare in the vicinity, whereas desert sage and bitterbrush underwent local extinction. Plant species typical of full-glacial vegetation, such as limber pine and mountain mahogany, may have persisted into the latest

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Wisconsin, perhaps growing as scattered relicts in the pinyon-juniper wood-
land. The Holocene development of vegetation in the Eleana Range included the
apparent demise of these full-glacial relicts, the reduction or extirpation of
some species that were not established until the latest Wisconsin, and the
immigration and increase of shrubs such as jointfir and saltbush (pl. 2).
All these changes took place in the context of a continuing pinyon-juniper
woodland (figs. 1 and 4).

CHRONOLOGY OF VEGETATION AND CLIMATIC CHANGE

Climate is assumed to have been the forcing mechanism behind vegetational
change evident in the Eleana Range-2 time series. Major alteration in local
climate had occurred before the deposition of the ER-2(2) macrofossil assem-
blage, dated at 11,700±85 yr B.P., and change began before 13,200 yr B.P., the
age of the last record of subalpine conifer woodland (pl. 2). Progressive
increases in the percentages of two relative thermophiles, dwarf-goldenbush
and bitterbrush, began early in the chronosequence, by about 15,900 yr B.P.
(pl. 3; table 2). Percentages of steppe-shrub macrofossils also began to
decrease millenia before the major alteration in plant-community composition.
Sorensen's index of similarity (Sorensen, 1948), expressing the percent flor-
istic similarity of each macrofossil assemblage to the modern flora, was used
as a measure of overall change in the Eleana Range-2 time series. Only
species of trees, shrubs, and succulents were used in calculating these co-
efficients (table 2). Progressive increase in similarity indices occurs
well before the major vegetational change (pl. 2). Local change toward an
increasingly xerophytic plant association and an effectively drier climate
probably was gradual and not pronounced at first, followed after as much as
4,000 years by a relatively abrupt change to vegetation dominated by
thermophiles.

Comparisons of the Eleana Range-2 time series with other paleoenviron-
mental records further define the timing of late Wisconsin climatic change.
Data from two sites that were studied in detail are particularly relevant.
Stratigraphic and surficial evidence of fluctuations in the level of Searles
Lake provides a well-controlled record of changes in effective runoff.
Presently a salt pan in the Mojave Desert of California (pl. 1), the site was
covered by a large lake during part of the late Wisconsin (Smith, 1979; Smith
and Street-Perrott, 1983). A second site, Rampart Cave in the Lower Grand
Canyon of Arizona (pl. 1), also has been a rich source of paleoenvironmental
information. Chronostratigraphic and packrat-midden studies at this site
provide additional detailed data on the timing of biotic change at the close
of the last ice age (Martin and others, 1961; Long and others, 1974; Phillips,
1977).

Variations in the level of Searles Lake were controlled, in part, by
variations in the influx from the Owens River, incorporating glacial melt-
waters from the Sierra Nevada and precipitation from catchment basins farther
north in the Owens River system (pl. 1) (Smith, 1979; Smith and Street-
Perrott, 1983). Marked lake-level fluctuations reflect overall changes in the
regional hydrologic budget, and these changes were climatically controlled.
Between about 21,000 and 16,000 yr B.P., Searles basin held an extensive
freshwater lake that overflowed much of the time into the adjacent Panamint

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Valley. Smith (1979, p. 111) and Mankiewicz (1975) cite stratigraphic and geochemical evidence for the initiation of dessication followed by a declining lake level after about 17,100 yr B.P. Applying a correction factor of -900 years to compensate for small carbon-14/carbon-12 ratios in bicarbonate-enriched waters (Peng and others, 1978) places the time of this change in effective moisture at about 16,200 yr B.P. (pl. 3). This drying trend changed Searles Lake to a shallow body of water by about 14,500 yr B.P., more than 1,000 years before the pronounced biotic change at the Eleana Range-2 site (pl. 3). The significance of the subsequent resurgence of Searles Lake levels, including two high stands between about 12,000 and 10,000 yr B.P., is discussed in a later section. It is sufficient to note at this point that evidence from the Eleana Range-2 site demonstrates that the climate during this second late Wisconsin episode of maximum lake levels was much different from the climate during the preceding episode of lake high stands (pl. 3).

Paleoenvironmental data from Rampart Cave comes from two sources: (1) Biogenic floor sediments that exceeded 2 m in depth; and (2) indurated packrat middens that occur in niches and side-rooms (Martin and others, 1961; Long and Martin, 1974; Phillips, 1977). Unfortunately, an arson-caused fire in 1976 destroyed a large part of the fossil record available in this cave. The late Wisconsin section of Rampart Cave sediments consisted of two units. Unit B was a seam of loose plant debris and fecal pellets gathered into the cave by packrats. It underlies Unit A, a 1-m thick accumulation of extinct ground sloth (*Nothrotheriops shastensis*) dung (pl. 3). The shift in depositional regime from packrat midden to sloth dung occurred between 14,800 and 13,140 yr B.P.

These data indicate that the reoccupation of Rampart Cave by ground sloths after the close of the full glacial episode, as well as their prior abandonment of the cave at the close of the middle Wisconsin (Martin and others, 1961; Long and others, 1974), was due to climatic change. Specifically, during much of the late Wisconsin, the ground sloth may have been restricted to latitudes farther south by low winter temperatures. We are aware of no radiocarbon dates on Shasta ground sloth that are within the Wisconsin-maximum time (about 21,000 to 15,000 yr B.P.). The final extirpation of ground sloths in the Rampart Cave area is dated to 10,980±115 yr B.P., the weighted average (Long and Rippeteau, 1974) of four ^{14}C dates on untrampled dung balls from the surface of Unit A (pl. 3) (Thompson and others, 1980). Radiocarbon analyses of megafaunal remains from stratified cave deposits throughout the Southwest demonstrate contemporaneous last-appearance dates of about 11,000 yr B.P. (Thompson and others, 1980; Spaulding and others, 1983). The timing of this event possibly was due to factors other than climate (Martin, 1973; Long and Martin, 1974; Spaulding, 1983c).

Detailed analysis of packrat middens from Rampart Cave (Phillips, 1977) reveals progressive change in plant-macrofossil assemblages that began by about 16,300 yr B.P. Analyses of principal components show a decrease in amplitude values derived from Rampart Cave plant-macrofossil assemblages during the last one-half of the late Wisconsin (pl. 3) (Phillips, 1977, pl. 2). Phillips' (1977, p. 84) vegetational stage III begins at about 14,000 yr B.P. By this time, xerophytic plant species had become common in the late-glacial woodlands of the Lower Grand Canyon (see also Mead and Phillips, 1981).

The vegetation and climate of the American Southwest during the Wisconsin glacial age were radically different from those of the present. Woodland occupied vast areas that now support sparse desert scrub, and large inland lakes occupied basins that have a particularly severe aspect under current climatic conditions. Studies of late Wisconsin environments of the Southwest have been summarized by Martin and Mehringer (1965), Mehringer (1977), Van Devender and Spaulding (1979), Smith and Street-Perrott (1983), and Spaulding and others (1983). Conflicting hypotheses exist concerning both the nature of late Wisconsin climate and the timing of the glacial-interglacial transition. No doubt disparities in interpretations of the climatic history of this region are, in part, due to the combining of evidence from different times within the late Wisconsin. The Eleana Range-2 time series, along with data from other packrat-midden sites in the southern Great Basin, and the Rampart Cave and Searles Lake chronosequences, can be used to eliminate some of these conflicting hypotheses. In addition, new paradigms provided by model simulations of late-glacial climate, based on the astronomic theory of climatic change (Schneider and Thompson, 1979; Imbrie and Imbrie, 1980; Kutzbach, 1983), provide an objective conceptual framework in which to assess the hypothesized proxy climatic record.

The glacial ages of the American West commonly are termed "pluvials," indicating that ice-age climates were characterized by substantial increases in precipitation relative to the meager quantities that fall in the desert today. Many paleoclimatic studies (such as Leopold, 1951, and Snyder and Langbein, 1962) have associated pluvial precipitation regimes with equable temperatures. As described in these reports, equability, a small seasonal range of temperatures, is the result of a postulated large decrease in summer tempertures ($-\Delta T_s \geq 7°$ C), combined with a smaller (but usually unspecified) decrease in winter temperatures. Consequent estimates of the decrease in average annual temperatures are conservative ($-\Delta T_a \leq 5°$ C) and, therefore, indicate a climate that was only moderately cooler than that of the present, or a mild-pluvial paleoclimatic hypothesis. Increased effective moisture in mild-pluvial models is the result of postulated substantial increases in precipitation (percent P = +50 to +100) (table 3). Although these paleoclimatic models usually incorporate increased equability, equability *per se* indicates neither mild nor frigid temperatures, only a small range between winter minima and summer maxima. The mild-pluvial hypothesis is the model most commonly invoked to account for effectively moist conditions during glacial ages in the desert West (table 3).

The cold-dry paleoclimatic reconstructions of Brakenridge (1978) and Galloway (1983) present a fundamentally different view. They suggest that the glacial-age climate of the American West was neither mild nor pluvial. Instead, substantial decreases of both winter and summer temperatures are proposed, resulting in a greater decrease in average annual temperatures ($-\Delta T_a \geq 8°$ C) than that indicated by mild-pluvial reconstructions (table 3). For the cold-dry models, increased effective moisture is achieved by substantial decrease of evaporation rates with little or no relative increase in precipitation (table 3).

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Table 3.--*Paleoclimatic reconstructions for the Wisconsin maximum in the American West*

[-ΔT$_a$, change, in degrees Celsius, in annual temperature; -ΔT$_s$, change, in degrees Celsius, in summer temperature; -ΔT$_w$, change, in degrees Celsius, in winter temperature; ΔP, change, in millimeters (mm), in annual precipitation; percent P, ΔP/modern P × 100]

Reference	Study area	Methods	-ΔT$_a$	-ΔT$_s$	-ΔT$_w$	ΔP (mm)	Percent P
Mild-pluvial reconstructions							
Antevs, 1952	Lake Lahontan, Nevada.	Hydrologic budgets	2.5 to 3	---	---	+80 to +160	+50 to +100
Broecker and Orr, 1958.	Lake Lahontan, Nevada.	Hydrologic budgets	5	---	---	+200	+80
Mifflin and Wheat, 1979.	Nevada, state-wide.	Hydrologic budgets	3	---	---	---	[1]+68
Snyder and Langbein, 1962.	Spring Valley, Nevada.	Hydrologic budgets	25	[3]7.2	---	+200	+67
Van Devender, 1973.	Western Arizona	Packrat middens	2 to 4	---	---	+120 to +220	---
Cold-dry reconstructions							
Brakenridge, 1978.	Montana to Arizona (lat 45°40' N. to 33°22' N.).	Relict cirques and cryogenic deposits.	48	[4]8	---	0	0
Galloway, 1970, 1983.	Southwestern United States.	Cyrogenic deposits	10 to 11	---	---	---	-20
Cold-pluvial reconstructions							
Leopold, 1951	Lake Estancia, New Mexico.	Hydrologic budgets and snowline changes.	6.6	9	2.8	+180 to +250	+50 to +70
Spaulding, 1983a	Southern Great Basin.	Packrat middens	6 to 7	7 to 8	56	---	+30 to +40

[1]Statewide average.
[2]Extrapolated by Morrison (1965) and Schumm (1965).
[3]Extrapolated by Schumm (1965) and Brakenridge (1978).
[4]Recalculated by Smith and Street-Perrott (1983).
[5]Minimum estimate.

Leopold (1951) and Spaulding (1983a) have proposed cold-pluvial paleo-
climatic models that differ from both the cold-dry and mild-pluvial models.
These reconstructions indicate greater relative decreases in average annual
temperatures ($-\Delta T_a$ = 6° to 7° C) and generally smaller percentage increase in
precipitation (percent P = +30 to +70 compared to percent P = +50 to +100)
(table 3) than those postulated in mild-pluvial reconstructions. However,
the temperature decreases are less than and the precipitation increases are
greater than those postulated for the cold-dry reconstructions. Although
estimates of summer and annual temperature decreases in the two cold-pluvial
models are comparable, Leopold's (1951) postulated modest winter-temperature
decrease ($-\Delta T_w$ = 2.8° C) would result in a more equable temperature regime
than that proposed by Spaulding (1983a) ($-\Delta T_w \geq$ 6° C).

Sufficient evidence is available from the southern Great Basin to test
these alternative models, provided that specified bounds are placed on the
terms "mild" and "cold" (referring to temperature) and the terms "pluvial"
and "dry" (referring to precipitation). The proposed bounds, summarized in
table 4, are tentative and arbitrary, but provide a basis for objective treat-
ment of models describing distinctly different temperature and precipitation
conditions. For this report, a pluvial paleoclimate is defined as one in
which the average annual precipitation was at least 20 percent greater than
present precipitation (percent P $\geq$ +20); a dry paleoclimate is one in which
average annual precipitation was less than 20 percent greater than present
precipitation (percent P < +20). Mild refers to a relative decrease in annual
temperatures not exceeding 5° C ($-\Delta T_a \leq$ 5° C), and cold is defined as a
paleoclimate with temperatures at least 6° C lower than those of the present
($-\Delta T_a \geq$ 6° C). Equability is a condition that commonly is characteristic of
the mild-pluvial reconstructions (Van Devender, 1976; Martin and Neuner, 1978;
Grayson, 1980; Lundelius and others, 1983).

Table 4.--*Definitions of paleoclimatic models*

Temperature regime	Precipitation regime	
	Pluvial	Dry
Mild	$-\Delta T_a \leq$ 5° C, percent P $\geq$ +20	---
Cold	$-\Delta T_a \geq$ 6° C, percent P $\geq$ +20	$-\Delta T_a \geq$ 6° C, percent P < +20

Recognizing that the southern Great Basin is only a part of the American
West, and that regional climatic differences during the late Wisconsin proba-
bly were no less than they are today, we wish to determine which of the alter-
native models (table 3) best accounts for the fossil record discussed here.
Of equal importance is a consideration of the contrast between paleoclimatic
records of the Wisconsin maximum (about 21,000 to 15,000 yr B.P.) and paleo-
climatic records of the terminal Wisconsin (about 12,000 to 10,000 yr B.P.).

CLIMATE OF THE WISCONSIN MAXIMUM

The Wisconsin maximum, or full glacial episode, was the culmination of the late Wisconsin stadial. This time was characterized by maximum development of North American continental ice sheets (Mickelson and others, 1983) and by maximum development of alpine glaciers and some pluvial lakes in the West (CLIMAP, 1976; Porter and others, 1983; Smith and Street-Perrott, 1983). The ages of local maxima vary somewhat about the date for the Wisconsin maximum of 18,000 yr B.P. agreed on by the CLIMAP Project members (CLIMAP, 1976; Gates, 1976). We suggest that 18,000±3,000 yr B.P. effectively encompasses the full glacial episode in the American West.

Plant species adapted to relatively cold, dry environments dominate the full-glacial Eleana Range-2 macrofossil assemblages. Limber pine is among the hardiest of western North America's conifers. At present, it commonly is the only tree in the high (in excess of 2,500 m) subalpine habitats of central Great Basin mountain ranges. Limber pine extends farther into the interior of the central Great Basin (Little, 1971), and to higher altitudes, than any other montane or subalpine conifer. This distributional behavior is evidence of its adaptation to particularly cold, xeric habitats. Steppe shrubs, such as sagebrush, horsebrush, and rabbitbrush commonly occur with limber pine. These shrubs also are the dominants throughout vast areas of steppe-desert in the West, areas characterized by a continental temperature regime, with severe winters and dry summers.

The fossil record from the Eleana Range and vicinity fails to provide evidence for plant species typical of the more mesic habitats in the region. Temperate-montane species, such as ponderosa pine, Douglas-fir (*Pseudotsuga menziesii*), and Gambel oak, are absent. This record is consistent with evidence from most other fossil localities in the southern Great Basin (Wells and Berger, 1967; Spaulding, 1981, 1983a; Thompson and Mead, 1982; Wells, 1983) but largely in conflict with previous vegetational reconstructions based on fossil-pollen evidence. Substantial pine-pollen frequencies in full-glacial sediments have been interpreted as indicating widespread forest or parkland vegetation composed of mesophytic conifers, such as ponderosa pine (Martin and Mehringer, 1965; Mehringer, 1967). Such upland vegetation would be consistent with a mild-pluvial climatic regime. However, recent studies demonstrate that, in the southern Great Basin, subalpine and woodland conifers expanded instead. Expansion of xerophytic pines (limber pine, Intermontane bristlecone pine, and, at lower altitudes, pinyon pine) could account for the large pine-pollen percentage in full-glacial sediments and indicate a climate somewhat colder and drier than that proposed for mild-pluvial reconstructions (table 3).

Several independent approaches were taken to estimate $-\Delta T$ for the Wisconsin maximum in the vicinity of the Eleana Range (Spaulding, 1983a). Calculations based on the minimum decrease in winter temperatures necessary to exclude thermophiles, such as creosote bush (*Larrea divaricata*), from the study area lead to a postulated $-\Delta T_w$ of at least 6° C. Estimates of summer-temperature decrease sufficient to account for the altitudinal depression of

limber pine and Utah juniper yield a $-\Delta T_s$ of 7° to 9° C. Sites with vegeta-
tion broadly analogous to the Eleana Range-2 site vegetation during the
Wisconsin maximum are typified by annual temperatures at least 5° C lower than
those estimated for the fossil site today (Spaulding, 1983a).

The Eleana Range-2 fossil record is consistent with a full-glacial
climate that was 6° to 7° C colder annually and was characterized by an in-
crease of more than 20 percent in annual precipitation, concentrated in the
winter months. Thus, a cold-pluvial model is indicated for the Wisconsin
maximum in the southern Great Basin. Although some increase in average annual
precipitation occurred (maximum 40 percent), the increase was less than pro-
posed in mild-pluvial climatic reconstructions (table 3). Drought-tolerant
plant species abound in the Wisconsin-maximum fossil record from the southern
Great Basin and vicinity (Mead and Phillips, 1981; Spaulding, 1981, 1983a;
Thompson and Mead, 1982; Spaulding and others, 1983), despite recent claims to
the contrary (Wells, 1979).

THE GLACIAL-INTERGLACIAL TRANSITION

The trend toward effectively drier conditions in the southern Great Basin
began about 16,000 yr B.P., as indicated by the time sequence from the Eleana
Range-2 site (pl. 2), studies of Searles Lake and Rampart Cave (pl. 3), and
stable-isotope studies of ground water in the Amargosa Desert (pl. 1)
(Claassen, 1983). In the southern Great Basin and vicinity, the major vegeta-
tional change at the end of the last ice age was time-transgressive (fig. 5).
Change from woodland to desert scrub and, at higher altitudes, from subalpine
conifer woodland to pinyon-juniper woodland, occurred first in the most xeric
habitats and at the lowest altitudes within a particular vegetational zone
(fig. 5) (Cole, 1982; Spaulding, 1983a). The oldest records of the transition
to postglacial vegetational conditions are from the driest sites, where
available moisture is limited.

The upward shift of the ecotones between desert scrub and thermophilous
pinyon-juniper woodland and, at high altitudes, between pinyon-juniper wood-
land and subalpine conifer woodland (fig. 5), probably began 2,000 to
4,000 years before the major vegetational change at the Eleana Range-2 site
(pl. 2). Major vegetational change at some sites occurred as early as about
16,000 yr B.P. (Spaulding, 1983b), long before the final extirpation of wood-
land from deserts at about 8,000 yr B.P. (Van Devender, 1977; Van Devender
and Spaulding, 1979). These data do not support the hypothesis of a "***
consistent, widespread, contemporaneous vegetational change throughout the
Southwest ***" at 11,000 yr B.P. (Van Devender and Spaulding, 1979, p. 709).
Rather, vegetational change was gradual, beginning at the close of the full
glacial episode and did not end at most sites until the middle Holocene.
Major change in vegetational type from, for example, woodland to desert scrub,
was abrupt at many sites but occurred at different times at different sites.
The differential timing of major vegetational change can be conceptualized as
the transgression of ecotones through given localities (fig. 5).

CLIMATE OF THE LATEST WISCONSIN AGE

Although evidence exists for increasing temperatures beginning at about 16,000 yr B.P., climatic conditions during the latest Wisconsin (12,000 to 10,000 yr B.P.), as well as the early Holocene (10,000 to 8,000 yr B.P.), remained effectively wetter than the present. Van Devender (1977) noted the presence of woodland in some desert habitats until the end of the early Holocene, and he suggested that the persistence of a glacial-age precipitation regime accounted for this phenomenon. Although effective moisture was undoubtedly greater than it is at the present in the desert West, temperature and precipitation regimes of the latest Wisconsin and early Holocene probably differed substantially from the regimes of both the full glacial episode and the present. Evidence for this difference occurs in the contrasts between chronosequences from the Eleana Range-2 site and Searles Lake. The Wisconsin-maximum high stand of Searles Lake was contemporaneous with subalpine conifer-steppe shrub vegetation at the Eleana Range-2 site (pl. 3). However, during the latest Wisconsin resurgence of Searles Lake (about 12,000 to 10,000 yr B.P.), thermophilous pinyon pine-Utah juniper woodland grew at the Eleana Range-2 site.

Increasing temperatures were the most important feature of climatic change during the last one-half of the late Wisconsin (17,000 to 10,000 yr B.P.). By about 11,700 yr B.P., average annual temperatures probably were within 2° C of present values, an increase of about 5° C in less than 5,000 years. Pinyon pine, Utah juniper, and prickly pear are thermophiles that had largely replaced subalpine and steppe species at the Eleana Range-2 site. The increase in seasonal temperature regimes may have been relatively greater for summer than for winter. This concept is based on probable correlations between variations in solar insolation during the winter and summer half-years and corresponding changes in winter and summer temperatures (Kutzbach and Otto-Bleisner, 1982; Kutzbach, 1983). By 11,000 yr B.P., incident solar radiation during the summer half-year was about 7 percent more than the present (Berger, 1978; Kutzbach, 1983). Consequent increases in summer temperatures would have resulted in a longer growing season (relative to the full glacial episode) and an altered precipitation regime, producing an environment at the Eleana Range-2 site that would not have been conducive to the persistence of subalpine vegetation.

Against this backdrop of a marked increase in thermal regimes during the last millenia of the late Wisconsin, the evidence must be considered for continued, and even increased, effective moisture. As ΔT_a approached zero, causing subalpine woodland to change to pinyon-juniper woodland at the Eleana Range-2 site, Searles Lake actually underwent a resurgence to full-stage conditions (pl. 3). Van Devender's (1977) proposition that a pluvial-type winter precipitation regime existed as late as about 8,000 yr B.P. attempted to explain the persistence of xerophytic woodland in desert habitats during this time. However, was precipitation during the terminal Wisconsin (12,000 to 10,000 yr B.P.) dominated exclusively by winter rains and snow? We suggest that both fossil evidence and theoretical justification exist for viewing the terminal Wisconsin as a time in which the precipitation of the southern Great Basin was greater during both the summer half-year and the winter half-year than it is today.

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Studies of terminal Wisconsin and early Holocene paleoclimates in other arid regions of the world provide valuable perspectives on the nature of latest-glacial climatic change in deserts. By application of computer-based climate-simulation models, Kutzbach (1981, 1983; Kutzbach and Otto-Bleisner, 1982) demonstrated the potential importance of subtropical maritime air as a precipitation source in low-latitude deserts during the last millenia of the late Wisconsin and the first millenium of the Holocene. Increased summer temperatures and greater contrast in thermal regimes are two predicted first-order climatic responses to solar-insolation conditions that occurred about 12,000 to 9,000 yr B.P. The "Milankovitch thermal maximum," accounting for terminal Wisconsin vegetational conditions in Alaska (Ritchie and others, 1983), can properly be conceived of as generating a suite of climatic phenomena; these phenomena are atmospheric responses to externally determined solar-radiation variations (Vernekar, 1972; Schneider and Thompson, 1979; Kutzbach, 1983). In deserts, high summer temperatures enhance advective flow of subtropical maritime air into the interior of the western United States by intensifying surface low pressure over very heated terrain (Kutzbach, 1981; Kutzbach and Otto-Bleisner, 1982; also see Hales, 1974). Moreover, increased warm-season temperatures could have resulted in more intense subtropical high-pressure systems, such as the Bermuda High, that today account for much of the summer precipitation in the monsoonal southwestern United States (Arizona, New Mexico, and West Texas) (Bryson and Lowry, 1955; Sellers and Hill, 1974; Van Devender and Wiseman, 1977).

We propose that increasing summer temperatures during the last 6,000 years of the late Wisconsin resulted in a radical alteration of seasonal-precipitation regimes in the southern Great Basin. Winter precipitation was dominant prior to 16,000 yr B.P., but by about 12,000 yr B.P., perhaps as much as one-half of the annual precipitation occurred during the summer half-year. Floristic evidence for this change in seasonality can be seen in the abundance of prickly-pear cactus in latest Wisconsin samples from the Eleana Range-2 packrat midden (pl. 2). Succulents are susceptible to summer drought (Shreve, 1964) and the present scarcity of prickly pear near the site is attributed to meager summer moisture. Summer rainfall in the southern Great Basin during the latest Wisconsin may have exceeded present quantities by more than the 50 percent suggested by Spaulding (1983a). This condition may have been coupled with winter temperatures lower than and winter precipitation greater than those of the present (Van Devender, 1977), and it may have resulted in the ground-water recharge period that is dated between about 15,000 and 9,000 yr B.P. (Claassen, 1983). Isotopic evidence indicates that the recharge occurred principally from snowmelt (Claassen, 1983). Thus, the postulated mild-pluvial climatic conditions of the terminal Wisconsin may have caused this recharge and not the cold-pluvial climate of the Wisconsin maximum. Only one radiocarbon date on Amargosa Desert ground water, from tuff or tuffaceous valley fill, dates to the Wisconsin maximum. The remaining 14 radiocarbon dates are younger than 15,000 yr B.P., although only 5 date to the interval 12,000 to 9,000 yr B.P. (Claassen, 1983, table 6).

Given the nature of the regional geography, advected maritime tropical air would have entered the southern Great Basin from the Gulf of California and Pacific Ocean to the south and southwest (pl. 1). If this paleoclimatic reconstruction approximates actual climatic conditions during the period of maximum summer and minimum winter insolation (about 12,000 to 9,000 yr B.P.)

(Schneider and Thompson, 1979, fig. 1), summer precipitation would have been even greater near the sources of moisture in the subtropical Sonoran Desert and periphery (pl. 1). Therefore, the biotic response to such a climate should be more apparent in the fossil record from that region. Such evidence exists in macrofossil assemblages from packrat-midden sites along the trough of the Colorado River (Spaulding, 1983b), a physiographic feature that today is important in channeling advected maritime air into the interior of the western United States (Huning, 1978). An elaborate fossil record from the Whipple Mountains (King and Van Devender, 1977; Van Devender, 1977; Mead and others, 1978) and the Rampart Cave area (pl. 1) (Phillips, 1977; Mead and Phillips, 1981) demonstrates the persistence of woodland into the early Holocene at low altitudes near the Colorado River. At these localities, the lower altitudinal limit of woodland was below 320 m altitude during the latest Wisconsin (12,000 to 10,000 yr B.P.). At the same time, farther north in the southern Great Basin, the lower limit of woodland was at altitudes above 1,000 m (fig. 5) (Spaulding, 1983b).

The anomalous occurrence of low-altitude woodland in the trough of the Colorado River, latest Wisconsin high stands at Searles Lake, and a period of ground-water recharge between about 15,000 and 9,000 yr B.P. in the Amargosa Desert can best be explained by a terminal Wisconsin pluvial climate. This mild-pluvial may have incorporated significantly enhanced summer precipitation relative to the present, some relative increase in winter precipitation, and winters that may have been about 2° C colder than present winter temperatures.

SUMMARY AND CONCLUSIONS

The fossil record from the Eleana Range indicates that marked climatic change occurred during the last 6,000 years of the late Wisconsin. Physical environments during the Wisconsin maximum at about 17,000 yr B.P. were much different from those of the latest Wisconsin (12,000 to 10,000 yr B.P.). Between about 16,000 and 12,000 yr B.P., a profound change in climate occurred, caused by increasing global temperatures. We concur with Galloway's (1983) assertion that fossil records from the last 3,000 to 4,000 years of the late Wisconsin cannot be used to infer earlier, full-glacial climatic conditions. The key to a better understanding of late-glacial paleoclimates is an appreciation of the fact that the last 6,000 to 7,000 years of the late Wisconsin was a time of considerable flux in climatic (and biotic) systems. No evidence exists for a pronounced, step-wise change in climate and vegetation at about 11,000 yr B.P., as has been claimed previously (Van Devender and Spaulding, 1979; Spaulding, 1983b). Rather, the transition from full-glacial to interglacial (Holocene) climatic conditions began by about 16,000 yr B.P. and probably continued well after 10,600 yr B.P., the date of the last sample in the Eleana Range-2 record. Postulated climatic conditions for the late Wisconsin are summarized in table 5.

The Wisconsin-maximum climate in the southern Great Basin was colder than most reconstructions have proposed. The fossil data are most consistent with a cold-pluvial climatic reconstruction. Plant species typical of current subalpine, xeric woodland, and steppe habitats abound in the fossil record of this period. Mesophytic trees and shrubs, and those sensitive to extreme cold, are largely absent from the full-glacial flora. A pluvial rainfall

Table 5.--*Summary of inferred Late Wisconsin climatic conditions*
[yr B.P., radiocarbon years before present (1950)]

Age	Temperature	Precipitation
Latest Wisconsin (mild-pluvial climate; 12,000-10,000 yr. B.P.)	•Continued warming. •Average annual temperature within 2° Celsius of present temperature by 11,700 yr B.P. •Winter temperature about 2° Celsius lower than present temperature. •Summer temperature about 2° Celsius higher than present temperature.	•Renewed pluvial conditions. •Average annual precipitation greater than 100 percent more than present precipitation and much more than precipitation during Wisconsin maximum. •Winter precipitation greater than present precipitation. •Summer precipitation greater than 50 percent more than present precipitation. •Summer precipitation perhaps as much as winter precipitation.
Transition time (15,000-12,000 yr B.P.)	•Trend toward warmer conditions.	•Trend toward drier conditions.
Wisconsin maximum (cold-pluvial climate; 21,000-15,000 yr B.P.)	•Average annual temperature greater than 6° Celsius lower than present temperature. •Winter temperature equal to or greater than 6° Celsius lower than present temperature. •Summer temperature 7° to 8° Celsius lower than present temperature.	•Average annual precipitation equaled 30 to 40 percent more than present annual precipitation. •Winter precipitation equaled 60 to 70 percent more than present precipitation. •Summer precipitation equalled 40 to 50 percent less than present precipitation and much less than winter precipitation.

regime is indicated, but it was a relatively dry pluvial. The full-glacial fossil flora of the southern Great Basin are inconsistent with a relative increase in average annual precipitation that was greater than 40 percent of present averages. Evidence for mild full-glacial winter temperatures is lacking. The seasonal range of temperatures at this time probably was nearly as extreme as that of the present. Winters were frigid, at least 6° C below winter temperatures of the present, and average annual temperatures were about 7° C lower than present.

In contrast to the Wisconsin maximum, average annual temperatures during the latest Wisconsin were about 5° C warmer. As a consequence, the precipitation regime underwent marked changes. Average annual precipitation at this time probably was greater than during the Wisconsin maximum. A terminal Wisconsin mild-pluvial climatic model, incorporating increased precipitation during both the winter and summer half-years, accounts for the distinctive nature of the regional macrofossil record at this time, the chronology of fluctuations in Searles Lake, and the late Wisconsin-early Holocene recharge in the Amargosa Desert. Woodland persisted in the lower Colorado River valley (Van Devender, 1977) for several millenia after the onset of desertification in the southern Great Basin (Spaulding, 1983b). This persistence is taken as evidence that maritime subtropical moisture, drawn northward into the interior of the western United States from the Gulf of California and the Pacific Ocean, was an important contributor to regional precipitation at this time. The impact of monsoonal precipitation at the Eleana Range-2 site is reflected in the increase of prickly-pear cactus during the terminal Wisconsin, as well as in the local demise of steppe shrubs by about 11,700 yr B.P.

Detailed reconstruction of latest Wisconsin paleoclimates and testing of the above hypotheses are major objectives of ongoing research in the southern Great Basin. However, sufficient data now exist to indicate that the paleoclimate is consistent with what would be predicted from the astronomic theory of climatic change. Summer solar-radiation maximum that occurred at about 11,000 yr B.P. in the northern hemisphere was the apparent cause of a suite of climatic phenomena (Bryson and Swain, 1981; Kutzbach, 1981; Ritchie and others, 1983). In the subtropical deserts of North Africa and the Indian subcontinent, model simulations and field evidence indicate that the consequent increase in monsoonal rainfall was as much as 80 percent of current amounts (Bryson and Swain, 1981; Kutzbach, 1981; Kutzbach and Otto-Bleisner, 1982). We suggest that a similar intensification of precipitation occurred before the end of the Wisconsin glacial age in the American Southwest. Van Devender (1977; Van Devender and Spaulding, 1979) pointed out that packrat middens dating to this period lack the summer-rain dependent perennials that typify the present monsoonal deserts. However, many of these species are frost-sensitive (Spaulding and others, 1983); lower winter temperatures may have prevented their northward expansion during the last few millenia of the late Wisconsin.

Although evidence exists for increased summer rains during the terminal Wisconsin (12,000 to 10,000 yr B.P.) and the early Holocene (10,000 to 8,000 yr B.P.), this evidence does not contradict Van Devender's (1977) assertion that increased winter precipitation was important in the climates of this time. Isotopic evidence from ground water in the Amargosa Desert, near the Eleana Range-2 site, indicates that the recharge there between about

15,000 and 9,000 yr B.P. was a consequence of increased influx from snowmelt (Claassen, 1983). Furthermore, the latest Wisconsin resurgence of Searles Lake probably cannot be attributed solely to increased summer rains; temperatures at this time presumably were too high to allow effective runoff during the summer half-year. Winter precipitation, therefore, must have been a major factor in the expansion of Searles Lake.

In conclusion, during the last one-half of the late Wisconsin, two distinct pluvial climates probably occurred in the southern Great Basin. The Wisconsin-maximum vegetational conditions indicate a cold-pluvial climatic reconstruction, differing from both mild-pluvial and cold-dry models (table 3). This full-glacial climate ended by about 16,000 yr B.P. and was both cooler and drier than the terminal Wisconsin pluvial climate. After about 16,000 yr B.P., temperatures (summer temperatures in particular) increased rapidly, resulting in radical alteration of seasonal precipitation regimes. By about 12,000 to 9,000 yr B.P., increased precipitation during both the winter and summer half-years may have caused average annual precipitation to be more than 100 percent greater than present precipitation. The hypothesis represents a significant change from an earlier estimate of 10 to 20 percent relative increase in annual precipitation for the latest Wisconsin compared to the present (Spaulding, 1983, table 10). The mild-pluvial climatic model now proposed best fits the fossil evidence for this period in the southern Great Basin. Because average annual temperatures approached those of the present at this time (Kutzbach and Otto-Bleisner, 1982), a substantial relative increase in precipitation is required to maintain high stands in southern Great Basin paleolakes and to support woodland in the sere lower Colorado River valley. The terminal Wisconsin pluvial climate probably had considerable effects on the hydrologic and biotic systems of the southern Great Basin. We anticipate that ongoing research will provide further refinements in reconstruction of this important climatic event.

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SUPPLEMENTAL DATA

Table 6.--*Packrat-midden sites, samples, and radiocarbon*
dates illustrated in plate 2

[yr B.P., radiocarbon years before present (1950)]

Site number	Reference	Site name	Sample number	Radiocarbon date (yr B.P)
27	Spaulding, 1981	Blue Diamond 3	A0100	15,040±650
28	do.	Blue Diamond 5	A0100	15,800±680
36	do.	Penthouse 3	A0200	21,210±440
36	do.	do.	A0300	20,880±370
37	do.	Penthouse 2	A0200	11,550±150
38	do.	Penthouse 1	A0100	8,100±120
38	do.	do.	A0200	19,400±300
39	do.	Willow Wash 1	A0100	22,420±720
40	do.	Willow Wash 2	A0100	4,125±90
41	do.	Willow Wash 4	A0100	21,350±420
41	do.	do.	B0100	9,820±110
41	do.	do.	C0200	19,020±750
41	do.	do.	D0100	17,700±740
42	do.	Flaherty Mesa 1	A0100	20,390±340
43	do.	Flaherty Mesa 2	A0100	18,790±280
44	Spaulding, 1977	Canyon Two	A0100	3,310±100
44	do.	do.	A0300	1,990±70
46	Spaulding, 1981	Deadman 1	A0100	17,420±250
46	do.	do.	A0200	16,800±250
46	do.	do.	A0400	18,680±280

Table 6.--*Packrat-midden sites, samples, and radiocarbon dates illustrated in plate 2*--Continued

Site number	Reference	Site name	Sample number	Radiocarbon date (yr B.P)
48	Spaulding, 1981	Desert View 1	A0400	5,210±95
52	do.	Spires 1	A0100	3,240±150
53	do.	Spires 2	A0100	18,800±130
53	do.	do.	A0200	16,200±350
55	do.	Basin Canyon 1	A0100	15,610±260
56	do.	Basin Canyon 2	A0100	9,365±320
56	do.	do.	B0100	19,200±580
57	do.	Basin Wash	B0100	3,520±100
58	Wells and Berger, 1967	Pintwater Cave	A0100	16,400±250
60	Spaulding, 1980	Eureka View 1	A0101	2,635±140
61	do.	Eureka View 2	B0100	1,580±140
61	do.	do.	C0100	3,930±180
62	do.	Eureka View 3	A0100	535±150
63	do.	Eureka View 4	A0100	5,595±210
63	do.	do.	A0200	6,795±190
63	do.	do.	A0300	5,435±220
64	do.	Eureka View 5	A0100	14,720±530
64	do.	do.	B0100	8,330±250
67	Spaulding, 1983	Last Chance 1	C0200	9,280±210
67	do.	do.	C0300	11,760±90

Table 6.--*Packrat-midden sites, samples, and radiocarbon*
dates illustrated in plate 2--Continued

Site number	Reference	Site name	Sample number	Radiocarbon date (yr B.P)
74	Spaulding, 1983	Owl Canyon 2	A0100	10,260±520
74	do.	do.	A0300	10,070±220
76	do.	Point of Rocks 1	B0100	9,840±150
76	do.	do.	B0300	11,680±650
77	do.	Point of Rocks 2	A0200	9,560±220
77	do.	do.	A0300	9,260±180
86	Spaulding, 1980	Horse Thief 1	A0100	10,690±280
91	Wells and Berger, 1967	Spotted Range 1	A0100	8,420±100
94	do.	Spotted Range 2	A0100	9,450±90
112	Spaulding, 1981	Sawmill Canyon	A0100	3,530±90
112	do.	do.	D0100	3,040±75
113	do.	South Crest	A0402	21,700±500
114	do	Hidden Forest 1	A0100	11,570±120
115	do.	Hidden Forest 3	A0100	1,600±120
115	do.	do.	B0100	820±100
115	do.	do.	C0100	10,060±130
115	do.	do.	C0200	11,940±140
115	do.	do.	C0300	11,860±160
121	Wells and Berger, 1967	Aysees Peak	A0100	9,320±300
122	do.	Mercury Ridge 1	A0100	9,000±250
123	do.	Mercury Ridge 2	A0100	7,800±150

Table 6.--*Packrat-midden sites, samples, and radiocarbon*
dates illustrated in plate 2--Continued

Site number	Reference	Site name	Sample number	Radiocarbon date (yr B.P)
124	Wells and Berger, 1967	Mercury Ridge 3	A0100	12,700±200
125	do.	Funeral Range 1	A0100	11,600±160
126	do.	Ranger Mountains 1	A0100	16,800±300
135	Spaulding, 1983	Specter Range 1	A0100	300±40
135	do.	do.	A0200	100±200
139	do.	Specter Range 2	A0201	18,740±90
139	do.	do.	A0202	19,280±350
139	do.	do.	A0300	23,270±80
141	do.	Specter Range 3	A0100	20,200±1,800
176	do.	Point of Rocks 3	A0100	14,810±400
200	Spaulding, 1981	Eyrie 2	A0100	2,920±90
200	do.	do.	A0200	1,960±65
201	do.	Eyrie 3	A0100	16,490±220
201	do.	do.	A0200	18,890±340
202	do.	Eyrie 5	A0300	19,750±450

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