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Groundwater Isotopic Evidence for Paleorecharge in U.S. High Plains Aquifers

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Compositions of δD, $\delta^{18}O$, ^{14}C, and ^{3}H are distinct in unconfined and confined groundwaters beneath the nonglaciated U.S. High Plains and reflect differences in average paleoclimatic conditions between the Holocene and the Middle to Late Wisconsinan. δD and $\delta^{18}O$ are more depleted in confined groundwaters than in unconfined groundwaters beneath the southern and central Plains but are more enriched in confined than in unconfined groundwaters beneath the northern Plains. The range in stable isotopic composition of meteoric water north-to-south across the High Plains was apparently smaller during the Middle to Late Wisconsinan than it has been during the Holocene. The greater δD and $\delta^{18}O$ depletion of confined groundwater beneath the southern and central Plains agrees with isotopic and noble gas data on other paleowaters that suggest Middle to Late Wisconsinan recharge temperatures were cooler than average Holocene temperatures by 5 to 8°C. The greater δD and $\delta^{18}O$ enrichment of confined groundwater beneath the northern High Plains adds to previously reported evidence for stable isotope enrichment in glacial age precipitation across the northern Plains to the eastern United States. Additional research is needed to evaluate how patterns and mechanisms of moisture transport and other climatic variables might explain isotopic composition of paleorecharge during the last glacial period. ©1995 University of Washington.

INTRODUCTION

Climatic change during the Late Quaternary has been inferred from the isotopic composition and noble gas content of groundwater in Africa (Srdoč *et al.*, 1982; Dray *et al.*, 1983; Andrews *et al.*, 1994), Europe (Rozanski, 1985; Moser *et al.*, 1989; Stute and Deak, 1989), North America (Perry *et al.*, 1982; Claassen, 1986; Phillips *et al.*, 1986; Plummer *et al.*, 1990; Stute *et al.*, 1992; Plummer, 1993), and Australia (Jacobson *et al.*, 1989). Comparison of isotopic compositions of dated groundwater in unconfined and confined aquifers thus can supplement paleoclimatic inferences based on pollen and other fossils, δD of tree cellulose, geomorphology, and geochemistry of sediments and cements. Such isotopic evidence is not found in all aquifer systems containing old groundwater. For an isotopic record of paleoclimate to be preserved and identified requires (1) isotopic composition of paleorecharge to have been distinct from Holocene isotopic composition, (2) paleoclimatic conditions to have persisted long enough for recharge of one isotopic composition to replace older groundwaters in both unconfined and deeper confined aquifers, and perhaps in an intervening confining layer, (3) the unconfined aquifer to have been flushed by Holocene recharge with an unique isotopic composition, and (4) transport or recharge rates to be too low during the past 11,000 to 18,000 years to have replaced the isotopically distinct paleowater in the confined aquifers with Holocene water. Just how a distinct isotopic composition of paleowater relates to specific climatic variables and moisture transport patterns can be ambiguous for many reasons (Gat, 1983; Grootes, 1993).

Paleorecharge temperature estimated from Wisconsinan-age groundwaters in west-central North America appears cooler than average Holocene temperature by 5 to 8°C (Claassen, 1986; Phillips *et al.*, 1986; Stute *et al.*, 1992). Dutton and Simpkins (1989) and Parkhurst *et al.* (1992) found that stable isotopic composition is stratified beneath the southern and central High Plains, respectively, with δD and $\delta^{18}O$ more depleted in the confined aquifers than in the unconfined aquifers. Dutton and Simpkins (1989) lacked groundwater age data and could not determine whether isotopic stratification is caused by an elevation effect or a climatic effect, because both are functions of temperature.

The purpose of this study is to interpret relative groundwater ages, distinguish vertical and lateral flow paths, and evaluate paleoclimatic implications of isotopic stratification of groundwater beneath the U.S. High Plains. Little is known about the age, isotopic composition, or hydrologic history of unconfined and confined groundwaters beneath the nonglaciated High Plains (Fig. 1). Nonhydrologic data on paleoclimatic conditions are sparse in this region (Galloway, 1983; Baker and Waln, 1985; Holliday, 1987). This study provides new ^{14}C and tritium data, draws together previous reports of isotopically distinct groundwaters, and extends the isotopic comparison of unconfined and confined aquifers to the

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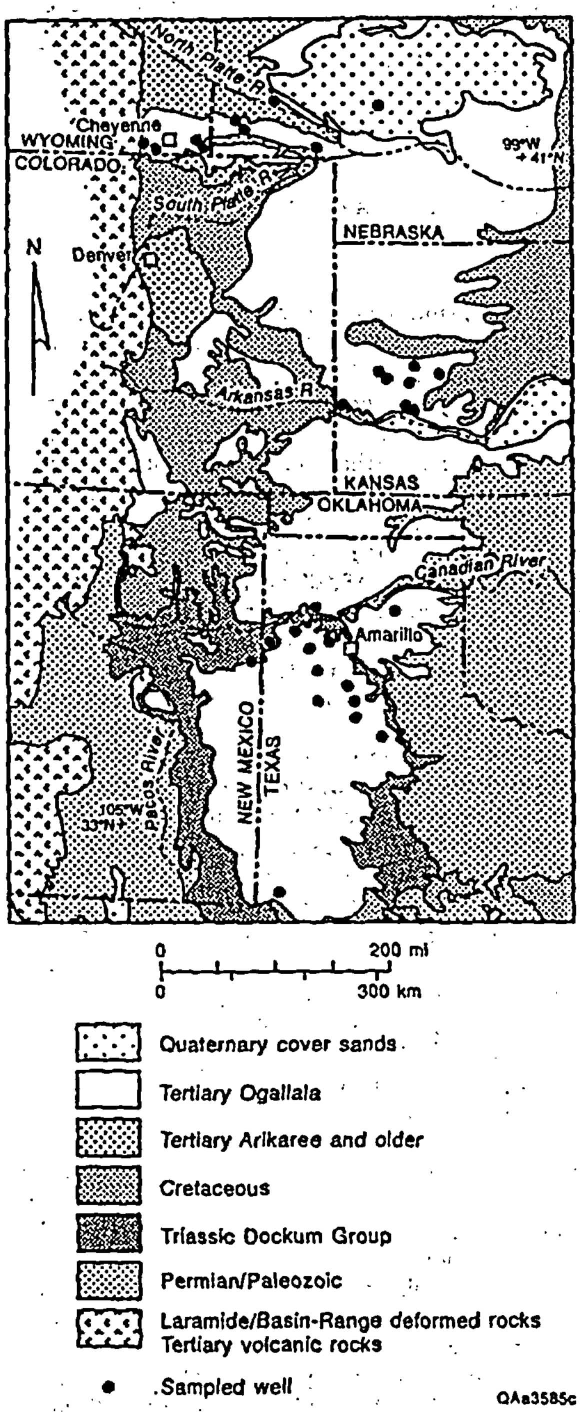

FIG. 1. Simplified geologic map of U.S. High Plains showing sampled wells in southern, central, and northern High Plains. Extent of the High Plains essentially the same as the Ogallala Formation (Osterkamp *et al.*, 1987). Gangplank refers to High Plains between the South Platte and North Platte Rivers. Quaternary surficial deposits not everywhere differentiated.

central and northern High Plains. The approach is to compare age and isotopic composition of groundwaters from paired wells in unconfined and confined aquifers. The 18 wells in confined aquifers from which samples were taken in this study (Figs. 1 and 2) probably represent 5–10% of available wells in confined aquifers in the three study areas. These wells were chosen to lie along presumed horizontal flow paths in the confined aquifers. To compare isotopic composition and groundwater age, wells in the unconfined aquifer were chosen for sampling that were close to the wells in the confined aquifer. The 14 sampled wells in the unconfined aquifer constitute fewer than 0.01% of drilled wells in the extensively developed High Plains aquifer.

HYDROGEOLOGIC SETTING

The High Plains in the midwestern United States (Fig. 1) is underlain by Tertiary and Quaternary alluvial fan, fluvial, lacustrine, and eolian deposits that formed when the Rocky Mountains eroded. Recharge to unconfined and confined aquifers in the Tertiary and Quaternary formations undoubtedly changed during the Pliocene to

		Northern	Central	Southern
Quat.		Undifferentiated alluvium	Mead Formation	Blackwater Draw Formation
Tertiary	P	Ogallala Group*	Ogallala Formation*	Ogallala Formation*
	M	Arikaree Group		
	O	White River Gp* (Brule / Chadron*)		
	E			
	P	Denver Fm.		
Cretaceous	U	Lance Fm.*		
		Fox Hills Fm.		
		Pierre Shale*	Pierre Shale	
		Niobrara Group	Colorado Group	
		Benton Group		
		Dakota Group	Dakota Sandstone*	Duck Creek Fm.
	L		Kiowa Shale	Kiamichi Fm.
			Cheyenne Sandstone	Edwards Fm.
				Comanche Peak Ls.
				Paluxy Sandstone
Jurassic	U	Morrison Fm.	Morrison Fm.	
	M	Sundance Fm.		
	L			
Triassic	U			Dockum Group*
	L	Chugwater Fm.		
Permian	U	upper Blaine Gypsum		Ochoan Series
			Guadalupian Series	Artesia Group
			Nippewalla Group	San Andres Fm.

* - Aquifers sampled in this study. QAa4527c

FIG. 2. Generalized correlation chart of stratigraphic units in northern, central, and southern High Plains. Based on Hills and Kottlowski (1983) and Kent *et al.* (1988).

Pleistocene when erosion of Pecos, Arkansas, and South Platte Rivers divided the southern and central High Plains from the Rocky Mountains (Gustavson and Finley, 1985; Osterkamp *et al.*, 1987). Erosion has not completely separated the northern High Plains from the Rocky Mountains in southeastern Wyoming and the Nebraska Panhandle: "At this place [the Gangplank], as nowhere else, you can step off the Great Plains directly onto a Rocky Mountain summit" (McPhee, 1986, p. 54). Hydrologic history of the northern High Plains might differ from that of the southern and central High Plains.

Groundwater flow in both unconfined and confined aquifers is generally from west to east parallel to the regional slope of ground surface (Gutentag *et al.*, 1984). The principal productive part of the unconfined High Plains aquifer lies in the lower part of the Ogallala Formation (Fig. 2). Recharge to the unconfined aquifer beneath the southern High Plains is focused through playa basins (Nativ and Smith, 1987; Osterkamp and Wood, 1987; Nativ and Riggio, 1989; Mullican *et al.*, 1994).

Groundwater resources in confined aquifers are sparsely developed except where the Ogallala is thin or absent. Beneath the southern Plains, the principal confined aquifer lies in sandstone in the lower part of the Triassic Dockum Group (Fig. 2), confined by upper Dockum mudstone (Dutton and Simpkins, 1989). A groundwater divide separates the confined aquifer from its outcrop in the Pecos and Canadian River valleys (Fig. 3). The hydrologic divide resulted from the mostly Pliocene excavation of the Pecos and Canadian River valleys (Gustavson and Finley, 1985).

Beneath the central Plains, the Cretaceous Dakota Sandstone composes the principal confined aquifer and is confined by Upper Cretaceous shales and carbonates (Belitz and Bredehoeft, 1988). Confined groundwater south of the Arkansas River flows eastward, but north of the Arkansas River it flows northeastward (Fig. 4). Less than 15% of recharge in the Colorado Piedmont flows under the Arkansas River to reach the confined part of the aquifer, where model results suggest that vertical

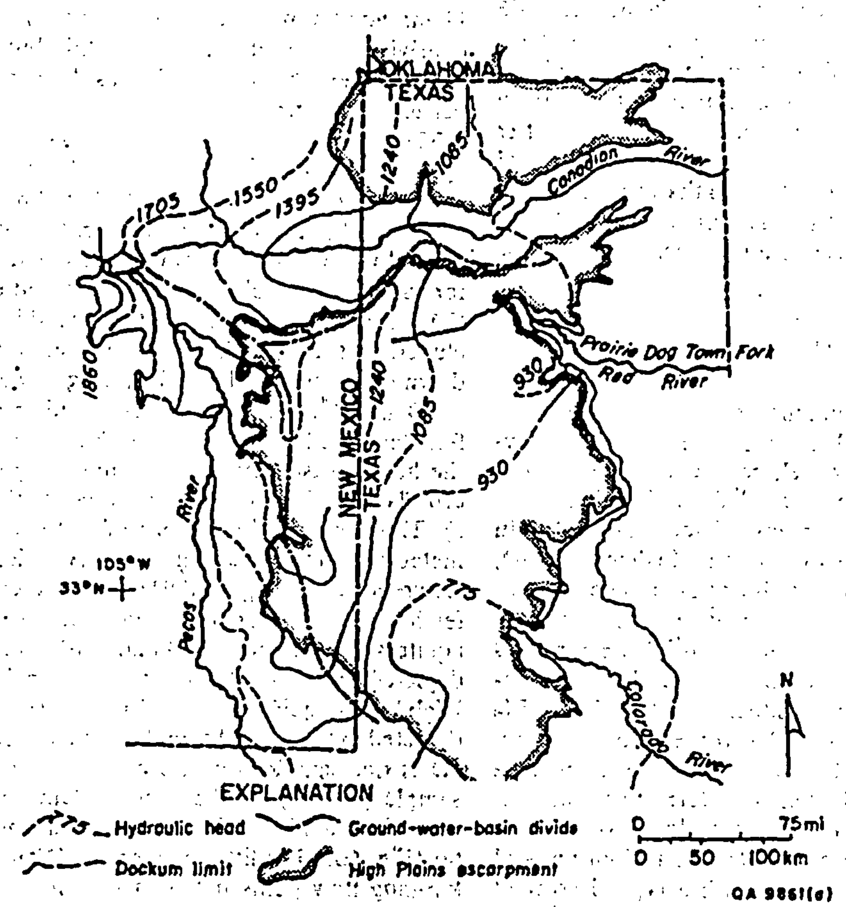

FIG. 3. Hydraulic head (meters above sea level) of groundwater in lower Dockum Group in southern Plains areas. Most of the 1056 water-level measurements used in contouring are from Dockum Group outcrop; only 77 are from the confined aquifer beneath the southern High Plains. Dashed-line contours signify areas of little well control. Symbol for High Plains escarpment is shaded on the High Plains side (see Fig. 1). Modified from Dutton and Simpkins (1989).

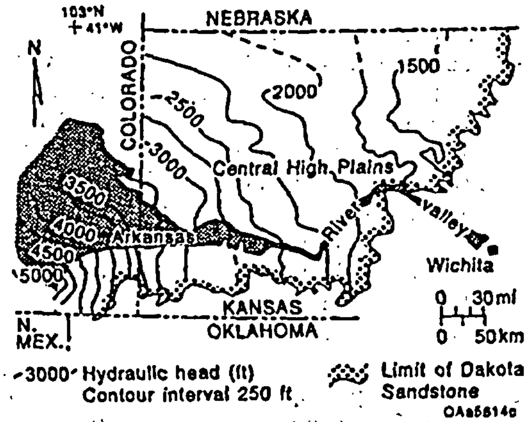

FIG. 4. Potentiometric surface of groundwater in Dakota Sandstone in central Plains area. Most of the ~570 water-level measurements are from the Arkansas River valley and its headwaters in the Colorado Piedmont (shaded area); less than 170 measurements are from the confined aquifer beneath the central High Plains. Modified from Macfarlane (1993).

leakage through confining beds accounts for at least 10% of total flow (Davisson *et al.*, 1993; Macfarlane, 1993).

Beneath the northern Plains, confined aquifers lie in an unconformable succession of Cretaceous to Tertiary rocks that underlie the Ogallala Formation (Swinehart *et al.*, 1985). Confined groundwater flows mostly eastward (Lowry and Christ, 1967) but flow paths probably bend north and south toward discharge areas in the North Platte and South Platte River valleys, where upward flow is possible.

METHODS

Most wells had been pumping nearly continuously for weeks before sampling; other wells were pumped for several hours to purge a few well-bore volumes of water. Waters to be analyzed for δD, $\delta^{18}O$, and tritium were filtered through 0.45-μm cartridge filters and sealed in sample bottles without other treatment. To determine δD, water samples were reacted with hot zinc metal to generate hydrogen gas; to determine $\delta^{18}O$, water samples were equilibrated with CO_2. δD and $\delta^{18}O$ are reported relative to standard mean ocean water (SMOW). Laboratory measurements are reproducible within 1.4 to 5‰ for δD and 0.09 to 0.3‰ for $\delta^{18}O$; these errors are consistent with differences between two laboratories of 5.2‰ in δD and 0.5‰ in $\delta^{18}O$. Tritium was determined on electrolytically enriched water samples by low-level proportional counting; results are reported as tritium units (1 TU is 1 ^3H atom/10^{18} H atoms) with a typical error of ±0.1 TU.

Dissolved inorganic carbon for ^{14}C and $\delta^{13}C$ analyses was collected by direct precipitation using a 30% ammonium hydroxide solution saturated with $SrCl_2$ (Hassan, 1982). The $SrCO_3$ slurry was decanted from carboys and later filtered and washed in the laboratory with negligible exposure to the atmosphere. The $SrCO_3$ powder was analyzed by liquid scintillation counting for ^{14}C and mass spectrometry for $\delta^{13}C$. ^{14}C is reported as uncorrected percentage modern ^{14}C activity (pmc) and $\delta^{13}C$ relative to Peedee belemnite (PDB).

RESULTS

Average ^{14}C differs significantly between unconfined (^{14}C of 55.3 ± 24.2 pmc) and confined groundwaters (^{14}C of <3.7 ± 3.6 pmc). There is no significant regional difference in ^{14}C between unconfined groundwaters beneath the southern High Plains (average of 40.3 pmc, standard deviation of ±13.8 pmc), central High Plains (66.7 ± 6.2 pmc), and northern High Plains (69.3 ± 34.1 pmc). Tritium was above background in half of the samples from the unconfined High Plains aquifer, and in these it ranged from 6.4 to 32.2 TU (Table 1). Nativ and Smith (1987) found that tritium in groundwater in the southern High Plains aquifer was between 4.2 and 73 TU where the overlying unsaturated zone was thin but between 0 and 2.8 TU where the water table was deep, reflecting greater vertical travel time.

There also is no significant regional difference in average ^{14}C of confined groundwaters beneath the southern, central, and northern Plains (4.1 ± 3.8, 5.5 ± 4.7, and <1.7 ± 1.1 pmc, respectively). Tritium was below background in all samples from the confined aquifers except for sample C5 (Table 1), which also had elevated nitrate (20.8 mg/liter $N-NO_3$). Data from this well were excluded on the suspicion that younger water was moving down to the confined aquifer owing to well construction.

δD and $\delta^{18}O$ from both unconfined and confined aquifers decrease from south to north across the High Plains, the heaviest isotopic composition being from the southern Plains and the lightest from the northern Plains (Fig. 5). This matches the pattern of isotopic composition of meteoric water in central North America (Taylor, 1974; Lawrence and Wright, 1991). Isotopic compositions differ locally, however, between unconfined and confined aquifers. In the southern Plains, δD averages 19‰ more depleted in the confined aquifer than in the overlying High Plains aquifer (Figs. 6a and 7). δD is 20 to 26‰ more depleted in two paired samples from the confined aquifer; δD is not significantly more depleted in a third paired sample (Fig. 6a). In the central Plains, δD is 12 to 28‰ more depleted in confined than in unconfined groundwater (Fig. 6b). The little change in δD and $\delta^{18}O$ along 75-km-long flow paths between sampled wells in the confined aquifer beneath the central Plains, however, is not statistically significant given the small sample size (Fig. 8). In the northern Plains, δD is essentially identical in the paired samples from confined and unconfined aquifers

TABLE 1
Tritium, ^{14}C, $\delta^{13}C$, δD, and $\delta^{18}O$ Isotopic Composition of High Plains Ground Waters

Sample ID	Latitude	Longitude	Unit[a]	Depth (m)	Sample date	Tritium (TU)	^{14}C (pmc)	$\delta^{13}C$ (‰)	δD (‰)	$\delta^{18}O$ (‰)	Well sets
					Southern High Plains						
S1a	35.538	102.289	Do	131.1	12/17/84	—	—	—	−66, −68	−9.1, −9.3	
S1b	35.538	102.289	Do	131.1	6/18/91	—	2.7 ± 0.2	−3.8	−68	−10.1	
S2a	35.121	103.021	Do	24.4	3/12/85	—	—	−4.2	−48, −46	−6.8, −6.9	
S2b	35.121	103.021	Do	24.4	6/19/91	—	1.5 ± 0.2	−7.1	−50	−7.48	
S3a	34.983	103.367	Do	304.8	3/18/85	—	—	−7.6	−73, −72	−10.7, −10.7	
S3b	34.983	103.367	Do	304.8	6/20/91	—	4.8 ± 0.2	−8.7	−59	−8.6, −8.4	
S4a	35.122	102.502	Do	265.8	3/12/85	—	—	−4.4	−76	−10.8	1
S4b	35.122	102.502	Do	265.8	6/19/91	0.00 ± 0.09	12.3 ± 0.2	−8.7	−70, −73	−11.7	1
S4c	35.122	102.502	Do	265.8	8/19/92	—	13.1 ± 0.2	−8.7	−67	−10.5	1
S5a	34.854	102.347	Do	291.1	11/27/84	—	—	—	−54, −55	−7.4, −7.4	3
S5b	34.854	102.347	Do	291.1	6/19/91	0.00 ± 0.09	3.1 ± 0.2	−7.9	−58	−7.96	3
S6a	34.734	101.864	Do	253.0	11/30/84	—	—	—	−74, −75	−10.7, −10.8	
S6b	34.734	101.864	Do	253.0	6/20/91	−0.14 ± 0.09	2 ± 0.1	−8.5	−78, −82	−11.7	
S7a	34.546	101.769	Do	252.4	3/13/85	—	—	−5.4	−73, −71	−10.4	2
S7b	34.546	101.769	Do	252.4	6/21/91	0.09 ± 0.09	4.9 ± 0.3	−9.6	−68	−9.5	2
S8a	32.244	102.647	Do	365.8	3/15/85	—	—	−9.6	−47, −49	−7.5, −7.3	
S8b	32.244	102.647	Do	365.8	6/21/91	—	<1	−9.7	−51, −47	−7.2, −7.2	
S9	34.854	102.347	Do	—	11/27/84	—	—	—	−67, −70	−9.6, −9.6	
S10	34.883	102.317	Do	246.0	4/27/84	—	—	—	−80	−11.2	
S11	34.904	102.317	Do	282.2	4/27/84	—	—	—	−83	−11.8	
S12	34.896	102.181	Do	224.6	4/27/84	—	—	—	−85	−12.7	
S13	34.833	102.344	Do	252.1	4/30/84	—	—	—	−71	−9.9	
S14	34.793	102.331	Do	290.2	4/27/84	—	—	—	−67	−9.1	
S15	34.558	101.774	Do	244.1	3/13/85	—	—	—	−75, −73	−10.6	
S16	35.015	101.752	Do	—	11/28/84	—	—	—	−55, −55	−7.5, −7.5	
S17	34.775	101.869	Do	244.8	4/24/84	—	—	—	−62	−9.1	
S18	34.439	101.881	Do	244.1	4/25/84	—	—	—	−50	−6.6	
S19	35.271	102.565	Og	76.2	8/18/92	0.01 ± 0.09	61.1 ± 0.4	−6.0	−38	−6.57	
S20	—	—	Og	109.4	8/18/92	−0.02 ± 0.09	32.4 ± 0.2	−6.9	−54	−8.24	
S21	35.515	100.963	Og	140.2	8/17/92	0.82 ± 0.09	20.8 ± 0.7	−7.3	−42	−6.73	
S22	34.449	102.295	Og	114.6	8/19/92	1.50 ± 0.09	46.5 ± 0.4	−7.6	−34	−6.39	
S23	33.918	101.375	Og	110.6	8/20/92	−0.04 ± 0.09	45.2 ± 0.5	−8.4	−41	−6.86	
S24	34.202	101.708	Og	98.5	8/20/92	5.39 ± 0.18	35.7 ± 0.3	−8.0	−31	−5.35	
					Central High Plains						
C1	38.045	102.011	Da	103.6	11/ 5/91	−0.11 ± 0.09	11.8 ± 0.2	−8.0	−88	−12.8	
C2	38.478	101.362	Da	320.0	11/ 6/91	0.02 ± 0.09	2.1 ± 0.1	−8.4	−92	−12.2	1
C3	37.947	100.973	Da	213.1	11/ 9/91	−0.02 ± 0.09	6.4 ± 0.2	−7.1	−87	−11.8	2
C4	38.471	100.399	Da	346.9	11/ 8/91	−0.03 ± 0.09	1.8 ± 0.2	−7.1	−96	−13	
C5	38.313	100.893	Da	304.8	11/ 7/91	5.56 ± 0.18	21.1 ± 0.2	−8.2	−52	−7.93	
C6	38.478	101.362	Og	48.2	11/ 6/91	6.44 ± 0.21	72.6 ± 0.9	−7.6	−64	−9.5, −9.6	1
C7	38.476	100.911	Og	65.2	11/ 8/91	0.92 ± 0.10	60.2 ± 0.8	−4.3	−64	−9.7, −9.5	
C8	37.921	100.987	Og	133.2	11/ 9/91	0.0 ± 0.09	67.4 ± 1.0	−5.8	−75	−10.4	2
					Northern High Plains						
N1	41.095	104.909	La	285.9	11/13/91	—	3.1 ± 0.5	−12.0	−107	−14.4	1
N2	41.155	104.116	La	277.4	11/13/91	—	2.2 ± 0.3	−7.4	−100	−13.2	2
N4	41.591	102.879	Ch	262.7	11/14/91	—	2.2 ± 0.3	−12.8	−98, −97	−13.7	
N3	41.582	103.674	P	237.7	11/16/91	—	<0.6	−8.9	−89, −92	−13	3
N5	41.093	102.523	WR	147.8	11/22/91	—	<0.6	−16	−98	−13.4	
N6	41.185	104.973	Og	57.3	11/12/91	−0.06 ± 0.09	107 ± 1	−12.3	−110, −108	−14.3	1
N7	41.177	104.067	Og	34.1	11/12/91	6.39 ± 0.23	88.3 ± 1	−11.1	−106	−14.2	2
N8	41.717	103.661	Og	98.8	11/15/91	0.01 ± 0.09	47.9 ± 1	−8.9	−114, 112	−15.2, −15.4	3
N9	—	—	Og	—	11/15/91	—	—	—	−99	−13.8	
N10	41.648	101.369	Og	185.3	11/23/91	32.2 ± 1.10	34 ± 0.5	−8.8	−78	−11	

[a] Do-Dockum, Og-Ogallala, Da-Dakota, La-Lance, Ch-Chadron, P-"transitional" Pierre, WR-White River.
— Not measured.

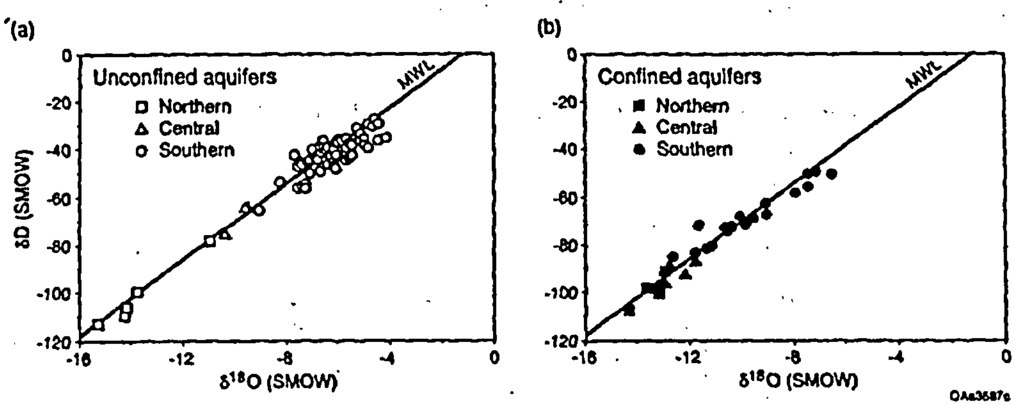

FIG. 5. δD and $\delta^{18}O$ in (a) unconfined and (b) confined ground water beneath the High Plains. Data given in concentration units. MWL is meteoric water line defined by $\delta D = 8\,\delta^{18}O + 10$ (Craig, 1961). Data from southern High Plains in (a) from Nativ and Smith (1987).

nearest the mountain front (set 1 of Fig. 6c; Fig. 9). Although the change in $\delta^{18}O$ is not statistically significant, δD becomes slightly more enriched eastward along the Gangplank—by 6‰ 75 km from the mountain front (set 2) and by 20 to 25‰ 110 km from the mountain front (set 3).

DISCUSSION

Key questions raised by these results are: (1) Does groundwater age constrain interpretation of the sources of confined groundwater? (2) What does isotopic stratification imply about paleoclimate? (3) What does residence time of isotopically distinct waters suggest about persistence of paleohydrologic or paleoclimatic conditions? (4) Does comparative hydrogeology of the southern, central, and northern High Plains reflect differences in hydrologic history?

Ages and Sources of Confined Groundwater

As discussed below, vertical recharge probably is superposed on lateral flow within the unconfined and confined aquifers. This means that there is mixing of groundwaters of varying ages along flow paths in both unconfined and confined aquifers. In addition, hydrodynamic dispersion results in mixing of successive pore volumes (Domenico and Robbins, 1985). Back-calculation of the effects of isotopic dilution and exchange for individual samples, therefore, would not uniquely estimate groundwater age. For the present purpose, ^{14}C and ^{3}H activities suffice to show that confined groundwater is significantly older than unconfined groundwater. Based on comparison with other studies (e.g., Phillips *et al.*, 1986), however, it is reasonable to assume that (1) age of unconfined groundwaters in the High Plains aquifer is generally less than 1000 yr and locally younger than 50 yr, and (2) isotopically distinct confined groundwaters are approximately 15,000 to 35,000 yr old and recharged during the last Pleistocene glaciation.

The assumed age of 15,000 to 35,000 yr is considerably younger than the (Pliocene) age of the hydrologic divide bounding the Pecos River valley (Fig. 3), suggesting that underflow from the Dockum Group outcrop in New Mexico was not the source of water in the confined aquifer beneath the southern High Plains. Vertical leakage, therefore, is the most likely source of water in the confined aquifer. Quantities of water moving in short but slow flow paths across regional confining layers can be large (Bredehoeft *et al.*, 1988). Climatic effects acting on the water that recharged the unconfined aquifer during the Middle to Late Wisconsinan, rather than an elevation effect acting across the Pecos Plains, must account for the isotopically light water now in the confined aquifers (Dutton and Simpkins, 1989). Senger and Fogg (1987) simulated a vertical travel time of approximately 310,000 yr downward across the confining layer beneath the southern High Plains. Their model results suggest that more than 90% of flow in the confined aquifer is derived from vertical leakage. Vertical hydraulic conductivity was set very low in model calibration (9×10^{-9} cm/sec). A 10- to 20-fold larger value of vertical hydraulic conductivity would allow even more leakage and give a vertical travel time of 15,000 to 35,000 yr. The decrease in ^{14}C activities along the northwest-to-southeast flow path in the confined aquifer (Figs. 3, 7) might reflect downward flow of groundwater across the confining layer and lateral transport in the confined aquifer. The greater ^{14}C activity in one sample at the east end of the transect (4.9 pmc, Fig. 7) might reflect more cross-formational flow near the High Plains escarpment (Senger and Fogg, 1987).

Davisson *et al.* (1993) also estimated that confined groundwater beneath the central High Plains is 15,000 to 40,000 yr old. Higher ^{14}C activities nearer the Arkansas River valley (6.4 and 11.8 pmc, Fig. 8) might reflect a shorter travel time and greater vertical leakage downward across the thin part of the confining layer. Lower

fined aquifer all along the Gangplank. The difference in ^{14}C activity between confined (2.2 to 3.1 pmc, Fig. 9) and unconfined groundwaters (88.3 to 107 pmc) nearest the mountain front reflects a long vertical travel time even in proximal-facies, coarse-grained Tertiary deposits.

Isotopic Stratification and Paleoclimate

The small variation in δD and $\delta^{18}O$ along flow paths within the confined aquifers does not appear to record short-term paleoclimatic changes. Instead, the variation most likely reflects vertical leakage and the paleoisotopic composition of the unconfined aquifer that was the source of water. Beneath the southern Plains, for example, the band of isotopically depleted confined groundwater coincides with the area where the confining layer is thinnest and probably allows the greatest leakage (Dutton and Simpkins, 1989). Beneath the central and northern Plains, the possible trends in δD in confined groundwater might reflect mixing of groundwater that was recharged in the Colorado Piedmont or water that was recharged near the mountain front, respectively, and water that leaked downward from the High Plains aquifer.

The fact that δD and $\delta^{18}O$ in confined groundwater are more enriched beneath the northern Plains but more depleted to the south, compared to groundwater in the unconfined High Plains aquifer, suggests that the gradient in isotopic composition of meteoric water across the northern to southern Plains was less during the Middle to Late Wisconsinan than during the Holocene (Fig. 5a). The δD isopleths of meteoric water (Taylor, 1974) appear to have been shifted farther north across the northern Plains and farther south across the southern Plains. Other confined groundwaters in the southwest United States have similarly depleted $\delta^{18}O$ as those from the southern and central

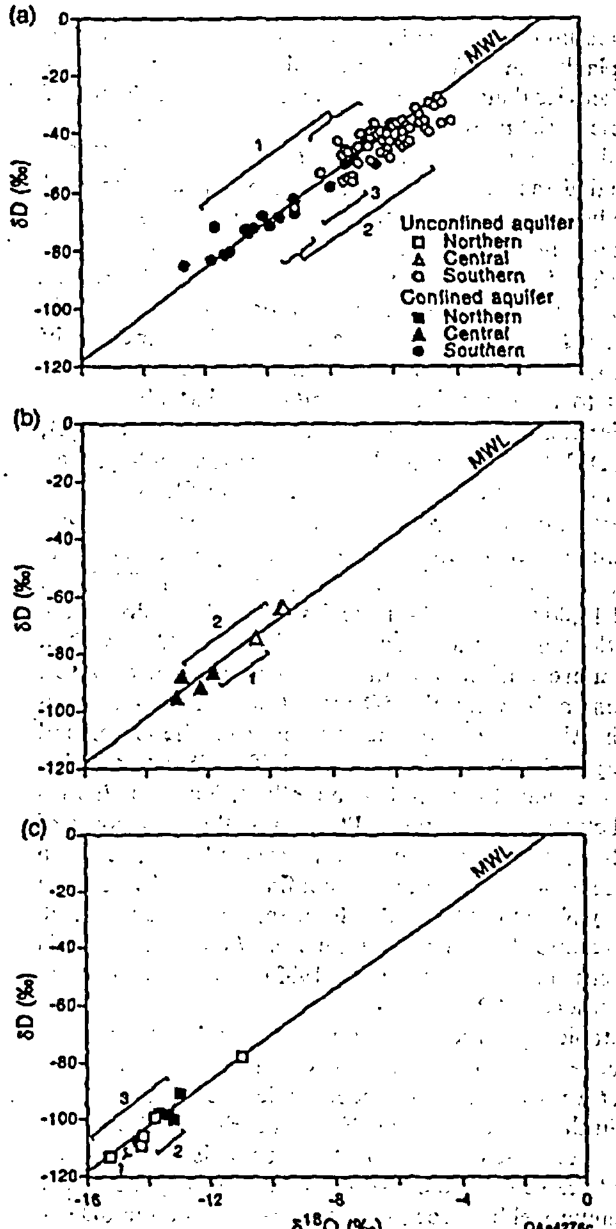

FIG. 6. δD and $\delta^{18}O$ in unconfined and confined groundwater from beneath (a) southern, (b) central, and (c) northern High Plains. MWL as defined in Figure 5. Numerals refer to paired wells (brackets). Data on unconfined aquifer in (a) from Nativ and Smith (1987), including their sample 59 (set 1 unconfined aquifer comparison), sample 53 (set 2), and sample 50 (set 3).

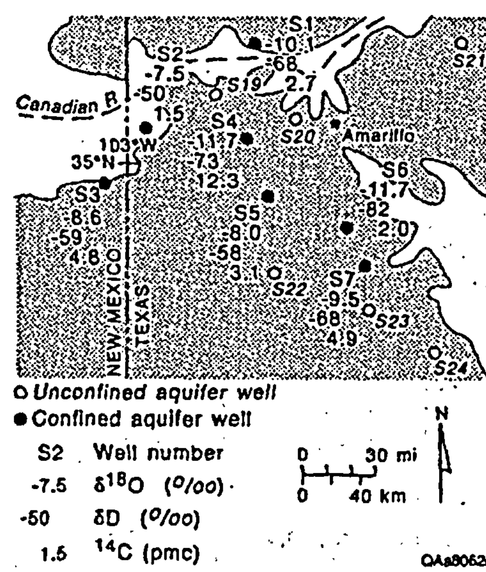

FIG. 7. Distribution of $\delta^{18}O$, δD, and ^{14}C is confined groundwater in Dockum Group beneath the southern High Plains (shaded area).

^{14}C activities farther to the northeast might reflect greater horizontal travel time (Fig. 4) and less vertical leakage across the thick part of the confining layer.

^{14}C activity in confined aquifers decreases from west to east beneath the northern High Plains (Fig. 9). These confined groundwaters were probably emplaced by both focused recharge at the western limit of the northern High Plains and by downward leakage from the uncon-

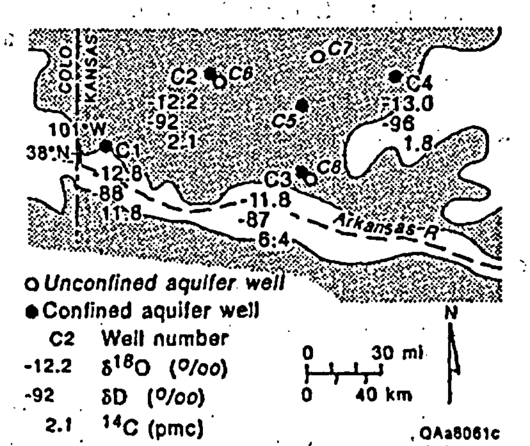

FIG. 8. Distribution of $\delta^{18}O$, δD, and ^{14}C in confined groundwater in Dakota Group beneath the central High Plains (shaded area).

Plains, and have been interpreted on the basis of stable isotopic composition and dissolved noble gas data as reflecting a Wisconsinan paleorecharge temperature 5 to 8°C cooler than average Holocene temperature (Claassen, 1986; Phillips *et al.*, 1986; Stute *et al.*, 1992). Some aspect of the last glaciation is the most likely explanation for the inferred narrow range in isotopic composition of meteoric water, including: (1) temperature difference between the moisture source and the precipitation area, (2) distance from ocean source(s), (3) jet stream and surface wind trajectories, (4) paleoisotopic composition of sources of moisture in Pacific and Atlantic Oceans and Gulf of Mexico, (5) differences in seasonal precipitation, and (6) an "amount effect," in which isotopic composition tends to be more depleted in large precipitation events (Dansgaard, 1964; Gat, 1983; Phillips *et al.*, 1986; Lawrence and White, 1991).

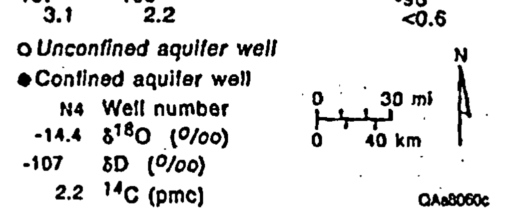

FIG. 9. Distribution of $\delta^{18}O$, δD, and ^{14}C in confined groundwater in confined aquifers beneath the northern High Plains (shaded area). Gangplank refers to High Plains between the South Platte and North Platte Rivers.

In addition to confined groundwaters from beneath the Gangplank, other Wisconsinan paleowaters are isotopically heavier than Holocene precipitation. The degree of isotopic enrichment possibly decreases from the eastern United States west to the northern Plains. Siegel (1991) concluded that a plume of dilute, isotopically enriched groundwater in the glaciated Plains of Iowa is a mixture of glacial meltwater that had a $\delta^{18}O$ of −20‰ and precipitation that had a $\delta^{18}O$ of −6 to −5‰. Such precipitation would have been 2 to 5‰ more enriched in $\delta^{18}O$ than local Holocene precipitation. Similar subglacially recharged groundwater in western New York was inferred to have received precipitation that also had a heavier-than-Holocene isotopic composition (Siegel, 1992; D. I. Siegel, personal communication, 1994). Plummer (1993) identified paleowater from the Floridan aquifer system in the southeastern Atlantic coastal plain with maximum $\delta^{18}O$ enrichment of 0.7 to 2.3‰ relative to local Holocene groundwater. Yapp and Epstein (1977) found δD of 14,000- to 22,000-yr-old fossil wood cellulose from near the limit of the Wisconsinan ice front that is 20 to 30‰ more enriched than Holocene meteoric water. Agreement between the δD record of dated wood cellulose in Alaska and isotopic records from the Greenland ice cap and Atlantic Ocean foraminifera (Epstein, 1993) gives credence to the cellulose method. Not all paleowaters between the northern Plains and the east coast of North America, however, are isotopically heavier than Holocene precipitation; other confined groundwaters are more depleted in $\delta^{18}O$ than Holocene recharge by 2‰ in north-central Wyoming (Plummer *et al.*, 1990), by 5‰ in northern Illinois (Perry *et al.*, 1982), and by 3 to 8‰ in Iowa (Siegel, 1991).

Isotopically enriched groundwaters in confined aquifers beneath the northern Plains probably do not simply reflect a warmer-than-Holocene paleorecharge temperature in the proglacial area. Yapp and Epstein (1977) hypothesized that evaporation from cold, isotopically enriched seawater of the North Atlantic during the glacial maximum yielded precipitation more enriched in δD and $\delta^{18}O$ than Holocene precipitation. One possible mechanism bringing moisture from the North Atlantic across the northern Plains is suggested by paleoclimate simulations (Kutzbach and Wright, 1985; COHMAP Members, 1988) that showed a weak anticyclonic circulation between two jet streams at the time of glacial maximum (~18,000 yr). Simulations suggest that anticyclonic circulation was strongest in winter months, but was overwhelmed by strong westerly air flow during summer months (Kutzbach and Wright, 1985; Kutzbach and Guetter, 1986). Alternatively, Siegel (1991, 1992) interpreted isotopically enriched groundwater in the glaciated areas as evidence of either a mild, proglacial climate or the influx of moist tropical air. Climatic simulations also indicate that southerly air flow from the Gulf of Mexico

across the continental interior was possible and that precipitation near the ice front might have been above average (J. E. Kutzbach, personal communication, 1994). These simulations may or may not support Siegel's (1991, 1992) interpretation. The confined aquifers beneath the nonglaciated southern and central High Plains, however, record isotopically depleted, not enriched, meteoric water between Siegel's (1991, 1992) study areas in Iowa and New York and source areas in the Gulf of Mexico and Pacific Ocean.

Persistence of Paleohydrologic Conditions

In general, the climatic imprint on the isotopic composition of confined groundwater must have been made by downward flow of water, either regionally, as in the southern High Plains, or locally, as perhaps at the west side of the northern High Plains. Whether water moved into the confined aquifers by vertical leakage or by horizontal transport from an aquifer outcrop, climatic conditions must have persisted long enough for Wisconsinan groundwater to replace older water in the confined aquifers. If confined groundwater beneath the High Plains is 15,000 to 35,000 yr old, then two to four pore volumes might have passed through the heterogeneous strata during the last glacial age from about 73,000 to 11,000 yr ago (Dansgaard *et al.*, 1969). Hydrodynamic dispersion increases the effective time and number of pore volumes needed for complete displacement of an older groundwater (Davidson and Airey, 1982; Domenico and Robbins, 1985) and makes the temporal records of ^{14}C and stable isotopic composition imprecise. The last pore volume to move into confined aquifers during the past 15,000 to 35,000 yr would not have completely displaced older groundwater. Since the end of the Pleistocene enough time has elapsed for recharge to replace Wisconsinan groundwater in the unconfined aquifers but not in the deeper, confined aquifers.

Model of Tertiary Recharge in Southern and Central Plains

The late Pleistocene to Holocene record of groundwater flow in the northern Plains might serve as an analog for Tertiary groundwater flow in the southern and central Plains. Before the erosion of the Pecos Plains and Colorado Piedmont (Fig. 1), both unconfined and confined aquifers in the southern and central Plains might have been recharged at high elevations near the Rocky Mountains. By the late Pliocene, recharge at the foothills of the southern Rocky Mountains was largely captured by the Pecos River, and downward vertical leakage became the major source of recharge to the confined aquifer beneath the southern High Plains. Beneath the central High Plains, flow of recharge from near the Rocky Mountains into the confined aquifer probably has been greatly reduced; 85% of the total recharge now is lost across the Colorado Piedmont and upper Arkansas River valley by discharge from local- and intermediate-scale groundwater flow systems (Davisson *et al.*, 1993; Macfarlane, 1993).

CONCLUSIONS

The isotopic composition of dated, confined groundwater can shed light on regional paleoclimate. The depleted δD and $\delta^{18}O$ composition of confined groundwater beneath the southern and central High Plains agrees with the isotopic composition of other Middle to Late Wisconsinan groundwaters from the southwestern United States and might indicate paleorecharge temperature 5 to 8°C cooler than average Holocene temperature in that region. Isotopic compositions of confined groundwater beneath the northern High Plains, Iowa, New York, and Florida and wood cellulose in Wisconsinan-dated trees are more enriched than Holocene meteoric waters. Isopleths of isotopic composition of meteoric water were shifted, decreasing the range in isotopic composition of paleorecharge across the U.S. High Plains relative to its Holocene range. Further study is needed to map the effect of moisture transport patterns and other climatic variables on the isotopic composition of paleorecharge water.

A distinct isotopic record being preserved in confined groundwater implies that hydrologic and climatic conditions persisted long enough for a water mass of one isotopic composition to replace an older water mass in unconfined and deeper confined aquifers. Because of aquifer dispersion, the isotopic record in confined groundwater beneath the High Plains averages climatic effects over several pore volumes recharged during 60,000 yr of the Middle to Late Wisconsinan. During the past 11,000 yr, recharge under a warm and dry climate in the High Plains has replaced unconfined groundwater but not deeper confined groundwater. Comparative hydrogeology suggests that groundwater flow beneath the northern High Plains might serve as an analog for the hydrology of the southern and central High Plains during the Tertiary.

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