

A Quantitative Model of Ground-Water Flow during Formation of Tabular Sandstone Uranium Deposits

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Abstract

Tabular sandstone uranium deposits constitute the largest uranium resource type in the United States. A major point of contention has been the nature and direction of the ground-water flow. This paper presents a quantitative simulation of regional ground-water flow during uranium deposition in the Westwater Canyon Member and Jackpile Sandstone Member of the Upper Jurassic Morrison Formation in the San Juan basin. Input for the four-layer finite-difference model was from thicknesses and sandstone/mudstone ratios in each layer from 1,600 drill holes and 100 measured sections and from paleotopographic reconstructions. Topographic slope, shoreline position, and density contrasts in the lake and pore fluids controlled the directions of flow and recharge-discharge areas. The most important results for uranium ore deposit formation are that regional ground water discharged throughout the basin, regional discharge was concentrated along the shore line or playa margin, flow was dominantly gravity driven, and compaction dewatering was negligible. A strong association is found between the tabular sandstone uranium deposits and major inferred zones of mixed local and regional ground-water discharge. The results of ground-water modeling favor the brine interface model over the lacustrine-humate model for uranium deposition.

Introduction

Problem of ground-water flow and uranium deposits

TABULAR sandstone uranium deposits constitute the largest uranium resource type in the United States. They account for approximately 65 percent of the total United States historical production and present reserves of uranium (Chenoweth and McLemore, 1989). They are known on every continent except Antarctica. Despite their importance and 40 years of intense study, the formation of these deposits remains controversial.

Tabular sandstone uranium deposits occur as tabular, originally subhorizontal bodies entirely within reduced fluvial sandstones of Late Silurian or younger age (Fischer, 1947, 1974; Finch, 1967; Granger, 1968; Hilpert, 1969; Turner-Peterson et al., 1986; Sanford, 1992). Throughout this paper, "reduced" and "oxidized" refer to iron unless otherwise noted. A fundamental disagreement between the various models of ore formation concerns the nature and direction of ground-water flow that led to uranium deposition. Some workers have proposed a brine interface model in which two fluids mixed along a density-stratified interface with flow in the upper fluid being gravity-driven downdip and subhorizontal (Fischer, 1947, 1974; Shawe, 1956, 1962, 1966, 1976; Granger et al., 1961; Granger, 1968; Wood, 1968; Melvin, 1976; Sanford, 1982, 1990a, 1990b, 1992; Granger and Santos, 1986; Northrop and Goldhaber, 1990; Wanty, 1990; Hansley and Spirakis, 1992). Others have advocated a lacustrine-

humate model in which a single fluid was driven by compaction downward and outward from overlying muds (Peterson, 1980; Turner-Peterson, 1985; Adams, 1986; Fishman and Turner-Peterson, 1986; Turner-Peterson and Fishman, 1986). The nature of the ground-water flow is crucial because it affects virtually every other aspect of deposit formation. The source of the uranium and associated constituents, the nature of the ore-forming reactions, the timing of ore deposition, and the consequent recommendations for exploration are all dependent on the fluid-flow paths. Previous models have suffered from the inherent ambiguities of reconstructing ancient ground-water flow from indirect evidence such as diagenetic alteration. No one has previously attempted to reconstruct the regional ground-water flow system directly based on fundamental hydrogeologic principles and quantitative modeling. Conceptual, qualitative models have been presented (Sanford, 1982, 1990a, 1990b, 1992). This paper is a start toward a complete quantitative hydrologic model for ground-water flow during tabular sandstone uranium deposit formation. It aims to discriminate between the brine-interface model and the lacustrine-humate model, and between two-solution and one-solution models. It further distinguishes between proposed mechanisms of fluid flow, such as gravity-driven flow and compaction.

A computer model of ground-water flow helps evaluate the relative importance of deep and shallow gravity-driven flow, compaction, and density-driven flow for a variety of sediment transmissivities, deposi-

tion rates, lake levels, and lake water densities. Many important controls on ground-water flow are evaluated for their influence on flow directions and rates. The calculations assume steady state flow and only partly incorporate density variations, so improved calculations are possible. As the following analysis shows, the results of the current calculations favor the brine-interface model.

Description of tabular sandstone uranium deposits

The general features of tabular sandstone uranium deposits worldwide are summarized elsewhere (Finch, 1967; Fischer, 1974; Adams and Saucier, 1981; Nash et al., 1981; Thamm et al., 1981; Finch and Davis, 1985; Granger and Finch, 1988; Shawe et al., 1991; Sanford, 1992). Typically, tabular sandstone uranium deposits are in lenticular, arkosic, fluvial sediments underlain by marine rocks, especially evaporites. The thin, subhorizontal, tabular-shaped bodies float in the sandstones with no apparent structural or lithologic control. Although generally peneconcordant (Finch, 1967), they undulate and locally cross the bedding at high angles. They are found in reduced sandstones within dominantly red-bed sequences of sandstone and mudstone. The deposits are commonly located at the transitional facies between dominantly fluvial sandstones and dominantly overbank and lacustrine mudstones. They tend to concentrate in syndepositional synclines. Organic matter, either as detrital woody fragments or structureless impregnations, is closely associated with uranium ore. The ages of the host rocks range from Late Silurian to Tertiary, the older age limit corresponding to the first appearance of woody land plants. Most of the deposits are thought to have formed shortly after deposition of their host sandstone, with some workers proposing mineralization almost immediately after deposition and others advocating mineralization after deposition of overlying units.

In the Colorado Plateau, the largest tabular sandstone uranium deposits are in sandstones of the Upper Triassic Chinle Formation and Upper Jurassic Morrison Formation. The ore-bearing sandstones commonly are overlain by overbank and lacustrine mudstones. Both sandstones and mudstones are tuffaceous. The major belts of deposits are spatially associated with thinning of underlying aquifers (Sanford, 1992). Most workers agree that mineralization in the Morrison Formation predated the major compaction during Late Cretaceous marine deposition. Thus, these uranium deposits probably formed during the latest Late Jurassic or earliest(?) Early Cretaceous. This paper focuses on deposits in the Grants uranium region of the San Juan basin. Here, tabular sandstone uranium deposits are mainly in the Westwater Canyon and Jackpile Sandstone Members of the Upper Jurassic Morrison Formation. Deposits in

the Salt Wash Member on the northwestern margin of the San Juan basin are included in the model area.

Geologic and hydrologic setting

The San Juan basin contains one of the principle uranium-producing regions in the United States (Figs. 1 and 2). The basin today consists of largely flat-lying sedimentary rocks of continental and marine origin up to 4,600 m (15,000 ft) thick (Peterson et al., 1965). Lower Paleozoic to Pennsylvanian sedimentary rocks are largely marine, whereas Permian to Jurassic sedimentary rocks are dominantly continental red beds. During the time of ore formation, the San Juan basin was at the foot of a semiarid to arid drainage area covering about 400,000 km² (155,000 mi²) and extending southwest (up-gradient) to the Mogollon highlands (Fig. 1; Sanford, 1982, 1990a and b). Eroding highlands merged northeastward into an aggrading alluvial plain dominated by sand, which merged into mud flat in the central part of the basin. Meteoric water entering the ground-water system in the highlands flowed north and east as deep gravity-driven flow toward the basin, where the ground water rose and discharged. Compaction of sediments may have yielded some additional deep ground water, but calculations below show this to be a relatively insignificant contribution. Infiltration from losing streams and direct precipitation in the alluvial plain and the Defiance uplift (Fig. 1) recharged shallow, gravity-driven flow. During the Late Jurassic to Early Cretaceous time of interest, braided streams deposited sand around the margins of the basin while mud flats occupied the center (Craig et al., 1955; Bell, 1986; Turner-Peterson, 1986; Turner and Fishman, 1991). With time, the mud-dominated facies transgressed south and west over the fluvial sands. The lower fluvial sands became the Westwater Canyon Member and the overlying muds became the Brushy Basin Member of the Morrison Formation. During part of the Brushy Basin deposition, an ephemeral, fresh to alkaline-saline, playa lake occupied the center of the basin (Craig et al., 1955; Keller, 1962; Sweet, 1981; Bell, 1983, 1986; Turner and Fishman, 1991). At the end of Brushy Basin deposition, fluvial sands of the Jackpile Member were deposited on top of the playa-lake mudstones in an active syncline (Moench and Schlee, 1967). Subsequent Early Cretaceous erosion removed some of the upper part of the Brushy Basin Member and beveled much of the Morrison Formation around the south and west margins of the basin. Marine transgressions in the Late Cretaceous led to major deposition, subsidence, and burial.

MODFLOW program

Paleoground-water flow in the San Juan basin was calculated using the U.S. Geological Survey com-

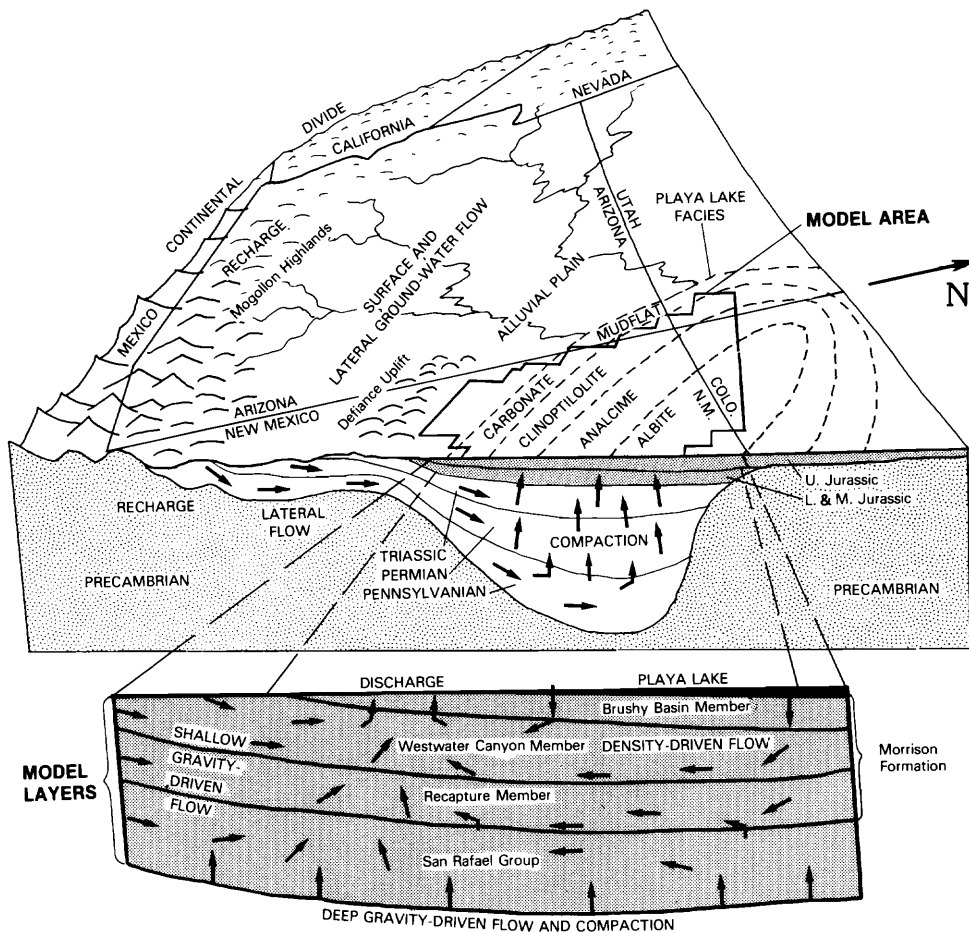


FIG. 1. Conceptual model of ground-water flow in the Late Jurassic to Early Cretaceous San Juan basin. Labeled model area includes the San Juan basin in northwestern New Mexico and part of the Paradox basin in southwestern Colorado and southeastern Utah. Sediment thicknesses from Peterson et al. (1965).

puter program, MODFLOW (McDonald and Harbaugh, 1988). The program has been intensely tested and is widely used in modern hydrologic studies. It simulates flow in three dimensions using a block-centered finite-difference method. The program solves the fundamental equation of ground-water flow (McDonald and Harbaugh, 1988):

$$\frac{\partial}{\partial x} \left(K_{xx} \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y} \left(K_{yy} \frac{\partial h}{\partial y} \right) + \frac{\partial}{\partial z} \left(K_{zz} \frac{\partial h}{\partial z} \right) - W = S_s \frac{\partial h}{\partial t}, \quad (1)$$

where x , y , and z are Cartesian coordinates aligned along the major axes of hydraulic conductivity, K_{xx} , K_{yy} , and K_{zz} ; h is the potentiometric head in meters (m); W is a volumetric flux per unit volume and represents sources and/or sinks of water in sec^{-1} ; S_s is the specific storage of the porous material (m^{-1}); and t is time (sec). This equation yields the distribution of po-

tentiometric or hydraulic head in the model area. In the present model, a steady state is assumed so that the $\partial h / \partial t$ term, and consequently the right hand side of the equation, equals zero. Future models will incorporate transient flow.

Once the head distribution is calculated, the flows within the model area and across the boundaries are calculated from Darcy's law:

$$q_x = -K_{xx} A_x \frac{\partial h}{\partial x}, \quad q_y = -K_{yy} A_y \frac{\partial h}{\partial y},$$

$$q_z = -K_{zz} A_z \frac{\partial h}{\partial z}, \quad (2)$$

where q_x , q_y , and q_z are Darcy flows in the x , y , and z directions, respectively, and A_x , A_y , and A_z are the cross sectional areas normal to the flow directions.

Description of the Computer Model

Input data to the MODFLOW program were chosen to simulate and test the effects of a wide range of

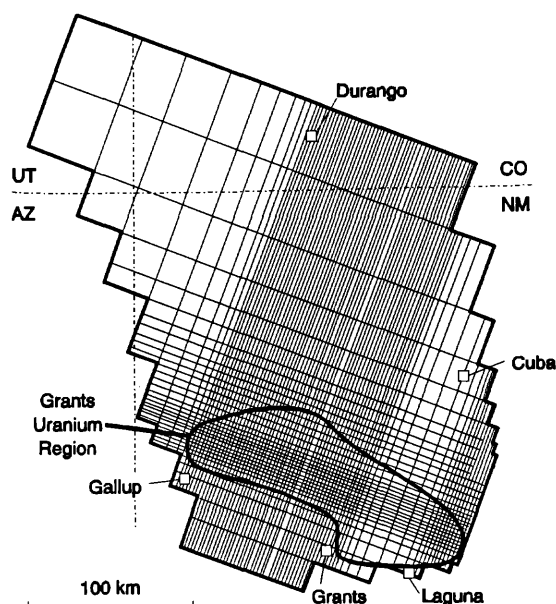


FIG. 2. Map showing active domain of finite-difference grid for the San Juan basin. The entire grid has 33 rows, 77 columns, and 4 layers. The active domain consists of 8,993 cells. The density of cells is roughly proportional to the density of drill holes.

basin configurations and environmental conditions. Within the constraints of the program, all reasonable variations of hydrodynamic properties and boundary conditions were tested, including hydraulic conductivity, deep regional flow, compaction, salinity, lake shoreline position, and erosion of basin margins.

Model area and hydrostratigraphic units

The model area is the San Juan basin where the Morrison Formation is present in outcrop and in the subsurface (Figs. 1 and 2). The model is 263 km (80 mi) by 311 km (95 mi) for a total surface area of 81,793 km² (31,600 mi²) and consists of 10,164 cells in four layers, 33 rows, and 77 columns (Figs. 2 and 3). The active domain of the model has a surface area of 56,517 km² (21,823 mi²) and consists of 8,993 cells. The layers (strata) from top to bottom are (1) the Brushy Basin and Jackpile Members of the Morrison Formation, (2) the Westwater Canyon Member of the Morrison Formation, including the Poison Canyon sandstone, (3) the Recapture and Salt Wash Members of the Morrison Formation, and (4) the San Rafael Group (Middle Jurassic). The rest of the drainage area up-gradient from the margins of the Morrison Formation and the formations below the San Rafael Group were treated as a single unit having recharge, compaction, and discharge to the model. The flow from this unit was incorporated into the boundary conditions of the model as described below. The

thickness and proportions of mudstone and sandstone in each layer were obtained from 1,600 drill holes and 100 measured sections (Kirk and Condon, 1986, 1989a-d; and unpublished data in U.S. Geological Survey data files). Geophysical logs for most holes consisted of four-track logs that included depth, natural gamma-ray, spontaneous potential, and single-point resistance logs. Gamma, resistivity, and locally, lithologic logs were used to pick the boundaries between members and formations and to estimate the thickness of mudstone and sandstone in each. The method of reducing the drill hole data to thicknesses of sandstone and mudstone is described in detail by Kirk and Condon (1986, 1989a). Because some of the data, from 23 oil and gas exploration companies, is proprietary, drill hole locations are not presented here. A sense of the drill hole density is given by the spacing of cells in the model (Fig. 2). Grid spacing was chosen so that one to three drill holes typically occupied each cell. The lengths and widths of cells consequently vary from 2 to 40 km with the smallest cells in the Grants uranium region. Contoured lithofacies and thickness maps have been published previously (Kirk and Condon, 1986, 1989a-d).

Calculation of hydrologic data

Hydraulic conductivity was calculated individually for each of the four layers at each drill hole and measured section. For each model run, a representative value of sandstone and mudstone hydraulic conductivity was selected from observed values compiled from diverse sources (Freeze and Cherry, 1979, p. 29; Ethridge et al., 1980). Representative hydraulic conductivities used in the model varied from 10⁻³ to 10⁻⁶ m/s and 10⁻⁶ to 10⁻¹² m/s for sandstone and

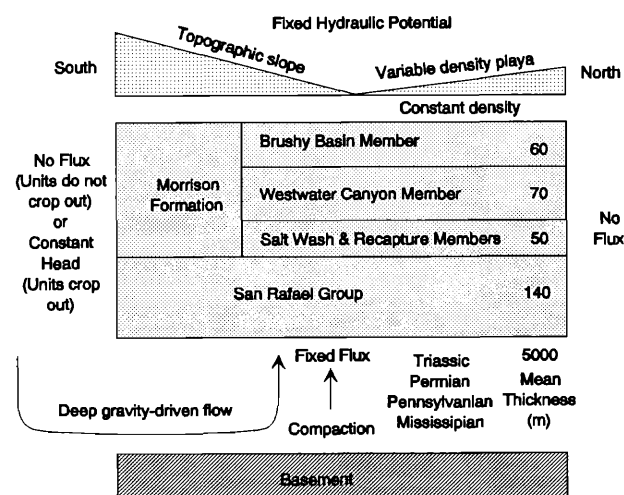


FIG. 3. Schematic cross section showing model layers and boundary conditions.

mudstone, respectively (Freeze and Cherry, 1979, p. 29; Ethridge et al., 1980). The observed thickness and sandstone/mudstone ratio were then used to calculate individual values of hydraulic conductivity at each location. These individual values are the thickness-weighted average of sandstone and mudstone hydraulic conductivities. The reader is referred to Kirk and Condon (1986, 1989a-d) for maps of sandstone and mudstone thicknesses used in the model. Average horizontal hydraulic conductivities were calculated from the arithmetic mean, and average vertical hydraulic conductivities from the hyperbolic mean (mean of reciprocals), of the mudstone and sandstone conductivities using individual mudstone and sandstone thicknesses for each layer in each drill hole. The variations in hydraulic conductivities used in the model (six orders of magnitude) are considered large enough to encompass the uncertainties due to variable grain size distribution, compaction, interconnectedness of sandstone lenses, cementation, and other factors. Hydraulic conductivities were further constrained, as discussed below, by means of flow rates for modern systems.

The calculated hydraulic conductivities for each layer in each drill hole were gridded by means of a least squares routine. The value at the center of each cell was calculated from the values for the nearest drill holes. A plane was fitted by least squares to the values of hydraulic conductivity at each of the nearest drill holes. The value at the center of each cell was then calculated by interpolation. A similar interpolation scheme yielded individual thicknesses of each layer in each cell.

Measured hydraulic conductivities vary by nine orders of magnitude. However, regionally averaged precipitation, recharge, and ground-water flow rates typically vary by less than two orders of magnitude, as discussed below. Thus, modern, measured, regionally averaged values of these properties can be used to constrain the input to the model.

Recharge rates constrain the model hydraulic conductivities and deep regional flow up through the base of the model, because the flow of ground water through the model cannot exceed the amount of water provided by recharge. The climate during deposition of the Morrison Formation was dry but with enough moisture to support fairly abundant vegetation mainly in riparian environments (Peterson and Turner-Peterson, 1987). Modern-day recharge has been estimated for climatically similar semiarid and arid parts of the United States and Australia. Typical precipitation in these areas is 300 ± 100 mm/yr. Only a small fraction of this infiltrates the ground deeply enough to recharge the ground-water system. Today, in the San Juan basin and southern high plains of New Mexico and Texas, recharge is estimated to have a range of 0.15 to 12.22 mm/yr and an average of $1.5 \pm$

1.0 mm/yr (Wood and Petraitis, 1984; Stone and McGurk, 1985; Stone, 1989; J.T. Dugan, U.S. Geological Survey, writ. commun., 1989). These figures are comparable to 1.0 mm/yr recharge calculated for the Great Artesian basin (Allison and Barnes, 1985). The recharge area for the Late Jurassic San Juan basin had an area of about $300,000 \pm 100,000$ km², whereas the discharge area was about $40,000 \pm 10,000$ km² as estimated previously (Sanford, 1990b, 1992) and verified in this paper. Thus the vertical discharge rate, ignoring compaction, would have been 5 to 10 times the recharge rate, i.e., about 15 ± 10 mm/yr (4.7×10^{-10} m/s). Constraining the hydraulic conductivity and upward flow through the bottom of the model so that discharge is within this range provides tight constraints on the overall water budget. By empirical testing, the model recharge is near the observed 1.5 mm/yr when the hydraulic conductivity for sandstone is about 10^{-3} and for mudstone about 10^{-8} m/s. These values are quite reasonable for the unconsolidated to partly consolidated sands and muds (Freeze and Cherry, 1979, p. 29; Ethridge et al., 1980) that formed the Jurassic sediments.

Flow rates in modern basins also provide tight constraints on the hydraulic conductivity. Darcian flow rates calculated from isotopic data and hydrologic modeling are commonly in the range 0.1 to 5.0 m/yr (3×10^{-9} – 1.5×10^{-7} m/s) (Belitz, 1986; Bentley et al., 1986; Phillips et al., 1986, 1989; Dowgiallo et al., 1990; Wei et al., 1990; Deming et al., 1992). The topographic slope of the model ranges from 5×10^{-4} in the alluvial plain to 5×10^{-5} in the playa. Hydraulic conductivities of 10^{-3} and 10^{-8} m/s for sandstone and mudstone, respectively, also yield observed flow rates of about 1 m/yr. Given the tight constraints on the flow rates, the hydraulic conductivities are probably within an order of magnitude of these values. The model flow rates are relatively insensitive to the sandstone hydraulic conductivity but are highly sensitive to the mudstone hydraulic conductivity, which demonstrates that the confining units control the flow rate.

Time frame of the model

The model simulates ground-water flow during deposition of the upper part of the Morrison Formation from early in the deposition of the Brushy Basin Member to the end of deposition of the Jackpile Member. This is the range of times most commonly proposed for ore formation. Fluxes, boundary heads, and pore water expulsion rate are time averaged for this period of deposition. Different thicknesses of the Brushy Basin Member were simulated by varying the vertical conductivity. Because flow probably was essentially vertical through the Brushy Basin Member, this approximation introduces no significant error.

Active and inactive cells

After the horizontal and vertical hydraulic conductivities were calculated for each cell, the active and inactive cells were selected. Cells that were beyond the limits of the drill hole data were made inactive. For each of the four layers (Figs. 1 and 3), the margins were smoothed, first, to eliminate cells with either three or four inactive adjacent cells, and second, to add new cells by interpolation where a cell without data was surrounded by three or four cells with data in the same layer. Then, for each vertical stack of cells, any cell that lacked an active cell above it was made inactive. This procedure resulted in a three-dimensional model with smooth edges and no isolated active or inactive cells. Finally, horizontal hydraulic conductivities were multiplied by the thicknesses of each unit in each cell to obtain transmissivities ($T_{xx} = T_{yy}$) for each cell. This process yielded individual values of total thickness, transmissivity, and vertical hydraulic conductivity for each active cell in the model. These data were formatted for input into MODFLOW.

Boundary conditions

Top surface—fixed head: The top surface of the model has a Dirichlet (fixed hydraulic head) boundary condition that simulates a water table sloping gently toward the basin center and a flat surface in the center of the basin. The flat surface represents equally the surface of a lake or a perfectly flat, dry playa. Because both conditions are characterized by a zero horizontal gradient, the two are hydrodynamically equivalent. Although this surface is referred to in this paper as the lake surface, the reader should remember that the results are the same for a flat, dry playa. The water table probably was some distance below the topographic surface in recharge areas but was near the surface in discharge areas, which constituted most of the model area.

The configuration of the top model surface, that is, the paleowater table in the San Juan basin, was reconstructed using the distributions of the lithologic and diagenetic facies of the Brushy Basin Member (Bell, 1983, 1986; Turner-Peterson, 1985, 1987; Turner and Fishman, 1991). The playa-lake facies are interpreted to represent different lake levels, and the boundaries between facies are interpreted as paleotopographic contours analogous to bathtub rings (Fig. 4). This interpretation assumes that the facies changes from alluvial facies to albite facies reflect increasing concentration of playa-lake water and that more concentrated playa-lake water ponded at successively lower elevations within the basin. Elevations were assigned starting arbitrarily with zero at the lowest point and rising toward the margins at slopes characteristic of large continental basins such

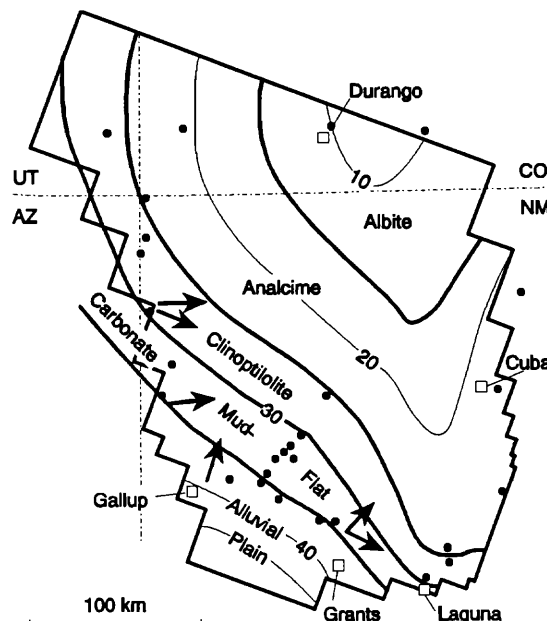


FIG. 4. Map showing reconstructed paleotopography in the San Juan basin during the Late Jurassic to Early Cretaceous. Paleotopography is based on lithologic and diagenetic facies of the Brushy Basin Member of the Morrison Formation (Bell, 1983, 1986; Turner-Peterson and Fishman, 1986; Turner and Fishman, 1991) and generalized current directions in sandstones (arrows, Turner-Peterson, 1986). Filled circles show data points. Paleotopography is contoured in meters (5-m contour interval).

as the Great Artesian basin, Australia (Habermehl, 1983), which has been suggested as an analogue to the Late Jurassic San Juan basin (Bell, 1983, 1986). Inferred slopes were typically 0.25 m/km in the alluvial plain and 0.10 m/km in the central playa. The resulting elevations ranged from zero in the northeast part of the basin to about 50 m in the southwest. To test the effect of fluctuations in lake level, the lake level was set for different runs at elevations from zero to 35 m. A zero lake level simulates times of extreme desiccation during Brushy Basin Member deposition when the playa was dry and times following the last appearance of the Brushy Basin lake. Lake levels up to 35 m simulate various depths of the lake during Brushy Basin deposition.

A first approximation was made for variable density of the ground water. The MODFLOW program assumes a constant density for the ground water. This is only an approximation to the real situation, since ground water in parts of the basin, especially the central playa, was probably saline or alkaline-saline (Bell, 1986; Turner and Fishman, 1991). Future models will incorporate variable density. However, as a first approximation, elevated salinity in the basin center was simulated by raising the hydraulic potential in the center of the playa an amount equivalent to the added potential of a column of brine that extends

down from the playa surface to half the thickness of the Brushy Basin Member. The value at half-thickness approximates the mean value through the whole Brushy Basin Member. In effect, this is equivalent to a hill in place of a flat playa. The maximum additional head due to salinity was calculated from the density of a brine saturated with halite (1.2 g/cm^3). Similar results would be obtained for brine saturated with gypsum or other soluble phases. Given a specific gravity of approximately 1.2 and a maximum half-thickness of the Brushy Basin Member of 60 m, the excess head due to salinity would be 20 percent of 60 m or 12 m. This roughly approximates the maximum additional weight of a column of salt water at a point in the middle of the Brushy Basin Member in the center of the basin. In different runs, the excess potential due to salinity was varied between zero, simulating a fresh-water lake or a flat playa, and 12 m, simulating the maximum density brine. When an excess head due to salinity was used, the excess head was set to zero at the shoreline and increased to the maximum at the basin center. This increase in excess head simulates the inferred increasing solute concentration of the playa-lake water toward the center of the basin.

Edges—no flow, fixed flow, and fixed head: The sides of the model have Neumann (fixed-flux) boundaries in most model runs. The Brushy Basin Member is beveled by erosion in the southern and southwestern parts of the basin where it is bordered by the Zuni and Defiance uplifts. Edges of units below the Brushy Basin Member may have been buried or exposed to recharge. Therefore, some model runs had a no-flux boundary along the southern and southwestern boundaries to simulate onlapping of the Brushy Basin Member over the area of the uplift, and other runs had a Dirichlet (fixed head) boundary to simulate exposure of the lower layers due to uplift and erosion. Most model runs simulated an intermediate case where the Westwater Canyon Member was exposed at the surface and underlying units did not crop out.

Bottom—fixed flow: The bottom surface was assigned various constant upward fluxes that simulate a range of deep gravity-driven flow and compaction. This flux varies from cell to cell according to the thickness of underlying sediments and varies in different model runs as a function of the rate of sedimentation for the Brushy Basin Member.

Deep gravity-driven flow was simulated by fixing the upward flux through the base of the model. This flow rate was estimated by calculating the water available from recharge in the highland areas surrounding the San Juan basin, as discussed below. As with other parameters, a range of values was used to evaluate the effect of natural variations and uncertainties in the data. The compaction rate was treated as a constant source (term "W" in eq. 1) and was estimated for layers within the model and for the rest of the

basin below the base of the model. Thicknesses and lithology of each unit were obtained from a variety of compilations (McKee, 1951; Baars, 1962; Jobin, 1962; Mallory, 1972). Each unit was classified as dominantly sandstone or mudstone. Then a value for the porosity change with depth was assigned based on the depth of burial during the time of Brushy Basin Member deposition and the porosity-depth curves for sandstone and mudstone of Hanor (1979). The variation of porosity with depth was then multiplied by the deposition rate for the Brushy Basin Member to obtain the rate of the pore-water expulsion (treated as injection rate). The deposition rate for the Brushy Basin Member was estimated from the present-day thickness and the estimated time for deposition (Kowallis et al., 1991). The effect of uncertainties was estimated by varying the deposition rate within a wide range.

Summary of input: In summary, measured and calculated values for hydrologic characteristics of modern ground-water systems tightly constrain the average values for the model as a whole. The ancient climate suggests precipitation of $300 \pm 100 \text{ mm/yr}$. Of this precipitation, about $1.5 \pm 1.0 \text{ mm/yr}$ infiltrated to recharge the ground-water system over an area of $400,000 \text{ km}^2$. The ground water flowed laterally at a rate of about 1.0 m/yr (range, $0.1\text{--}5.0 \text{ m/yr}$). Hydraulic conductivities were about 10^{-3} for sandstone and 10^{-8} m/s for mudstone, each with an uncertainty of an order of magnitude. Gravity-driven flow discharged in the basin at a rate of $15 \pm 10 \text{ mm/yr}$ over an area of $40,000 \text{ km}^2$. Given these overall constraints, the model then predicts the details of flow within the region of interest.

Results

The input parameters can be divided into two types, those that affect the position of recharge and discharge areas, and those that do not. Because the goal of the study has been to explain the location of uranium deposits, the amounts of flow are less important than the directions of flow. By far the major influence on ground-water flow directions is the configuration of the water table, or more precisely, the upper potentiometric surface. This surface was varied in two ways, by changing the size of the lake or playa surface and by changing the excess potential due to salt in the lake. The deep regional flow through the base of the Morrison Formation generally only affected the total flow through the model, but in some circumstances caused the flow to be reversed. Changing transmissivity, vertical hydraulic conductivity, and basin margin boundary conditions affected flow rates but not the locations of recharge and discharge.

Lake shoreline or playa margin position

The lake (or flat, dry playa) surface was set at selected levels from zero to 35 m above an arbitrary

reference level (Fig. 5). This variation represents both uncertainties in the actual lake level and real variations on the scale of seasons, years, and longer

period cycles. The lowest level is a close representation of (1) the time at the end of Westwater Canyon deposition before the Brushy Basin lake had formed,

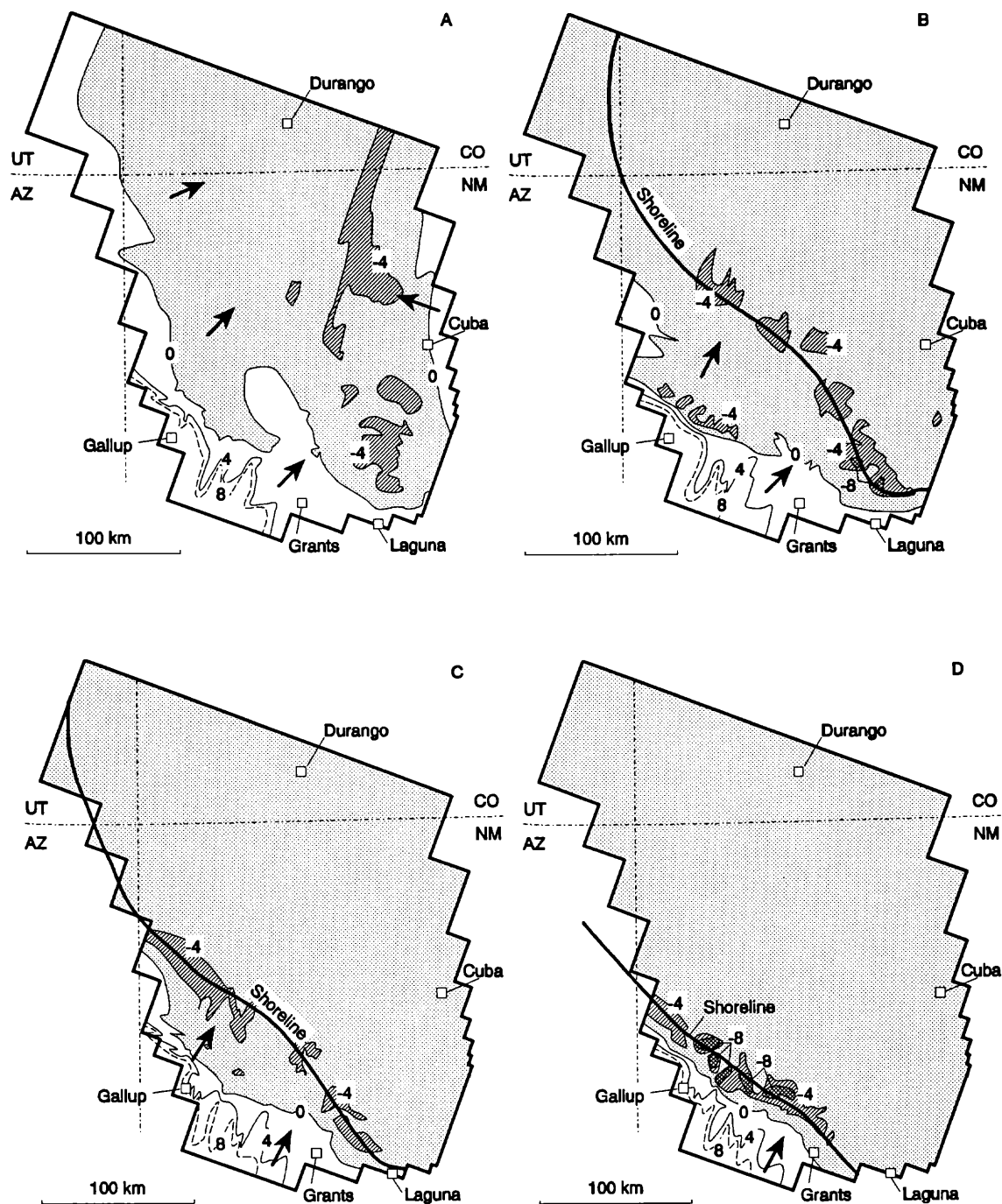


FIG. 5. Maps showing effects of changes in the position of lake shoreline on ground-water flow. Conventions for Figures 5 to 10 are the same: areas of discharge are shaded; different shadings indicate areas enclosed by the labeled contours, areas of recharge are white, contours show discharge (negative) and recharge (positive) in units of $m/s \times 10^{10}$; and arrows show directions of ground-water flow in the Westwater Canyon Member. A. No lake present at beginning and end of deposition of the Brushy Basin Member. B, C, and D. Shoreline at 25, 30, and 35 m, respectively.

(2) times of extreme desiccation during Brushy Basin deposition, and (3) the time of Jackpile deposition after the Brushy Basin lake had retreated. The highest stand, at 35 m, represents the greatest extent of the lake, when it reached the edge of the alluvial plain. Intermediate stands correspond to the lake shore at the different facies boundaries.

Figures showing recharge and discharge (Figs. 5–10) are contoured for vertical flux of ground water through the Brushy Basin Member. Recharge is indicated by positive values of flux and discharge by negative values. Vertical flux is expressed in units of 10^{-10} m/s. Shaded areas indicate discharge with greater shading density for higher discharge rates. The position of the shoreline in each model run is shown by a heavy solid line.

The following results have tremendous implications for the formation of tabular sandstone uranium deposits, as discussed below. In all model runs, most of the San Juan basin is an area of discharge (Fig. 5). Recharge is restricted to the margins of the San Juan basin. In the Grants uranium region, the hinge line separating discharge and recharge tends to be located where the alluvial plain merges into mud flat. This is where the topographic slope flattens out over a relatively short distance. Clearly, the topographic slope is the main control on the location of the hinge line and areas of recharge and discharge.

Local variations in the position of the hinge line and in the contours of equal discharge are controlled by slight variations in topography and permeability. For example, in the model with no lake (Fig. 5A), the hinge line shows a prominent northwest-pointing finger. This results from a slightly more gradual change in slope in the area of the finger than in the area just to the west. As an example of local variation due to permeability, the areas enclosed by the -4 contour are broken into islands having irregular boundaries rather than being combined into a single large area having a smooth boundary (Fig. 5A). Permeability variations thus affect the local recharge and discharge areas but are much less influential than topographic variations.

Comparison of four model runs having different lake levels reveals a systematic shift in the area of maximum discharge (Fig. 5A–D). The zone of maximum discharge rate occurs in the lowest part of the basin when a lake is absent, and at the shoreline when a lake is present. The position of the shoreline moves outward as the water level rises. When the lake is low, discharge is more dispersed over the whole area of discharge; when the lake is high, discharge is more focused or concentrated and the maximum upward flow is greater. The hinge line, or boundary between recharge and discharge, only moves slightly up-gradient to the southwest with a higher lake level.

The results show that variations in topography and

lake level are by far the most important controls on recharge and discharge areas. Next in importance is the effect of salinity and the excess hydraulic potential it adds.

Salinity and excess potential

The effect of alkaline-saline playa-lake water is simulated by increasing the hydraulic potential or head in the middle of the Brushy Basin Member by an amount equal to the water depth plus one-half the thickness of the Brushy Basin Member for water having 26 percent salt by weight. Because such a concentration is near halite saturation, this is probably the maximum excess head due to solutes. The maximum excess head in the center of the basin is calculated to be approximately 12 m.

This increased head in the center of the playa drives the flow downward and outward (Fig. 6). The model run illustrated in Figure 6 is identical to that of Figure 5B except for the increase in head in Figure 6. Horizontal flux is generally perpendicular to the contours of excess head. The playa-lake water discharges at the shoreline where it joins and mixes with deep regional ground water moving upward from the southwest.

Although the MODFLOW program assumes a uniform fluid density, and this approximation of the effect of salinity is admittedly crude, it does reproduce

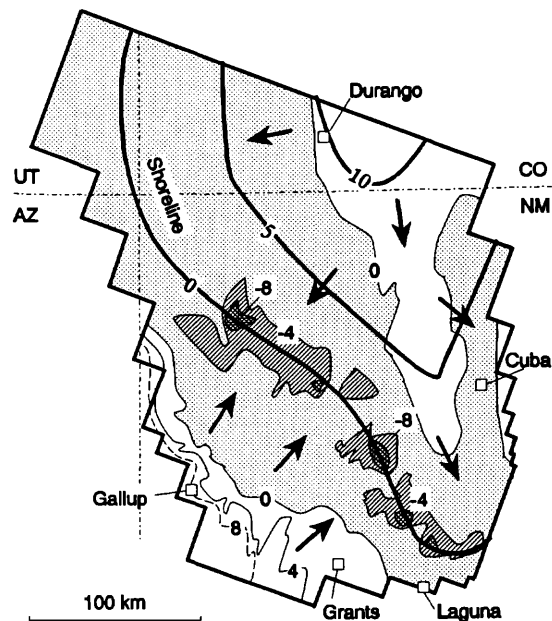


FIG. 6. Map showing effect of excess hydraulic potential due to salinity of ground-water flow. The excess potential in this run is 12 m at the deepest part of the basin at a depth of one-half the thickness of the Brushy Basin Member. Input conditions are the same as in Figure 5B, except for excess hydraulic potential, which is absent in Figure 5B.

the essential features of an alkaline-saline playa lake. As verified by more sophisticated variable-density models and by field observations, ground-water flow is downward in the center of the playa, outward toward the playa margins, and upward at the playa margin (Duffy and Al-Hassan, 1988; Macumber, 1992). Thus a largely closed convection cell is established beneath the playa lake, and mixing occurs at the playa margin near the spring line or zone of maximum discharge. An interface is present between the downdip flowing fresh ground water and the outward-moving alkaline-saline ground water from the playa. The main difference between constant- and variable-density models is that, when two compositionally distinct fluids of equal density meet, the interface is near vertical, whereas when fluids of different density meet, the interface is subhorizontal and extends up dip under the fresh-water lens.

Whereas the analogous modern systems (Duffy and Al-Hassan, 1988; Macumber, 1992) have playa-lake water mixing with shallow, dilute ground water, the present model for the San Juan basin suggests that playa-lake water mixes with deep, regional ground water. This difference arises from the scale of the flow systems. The San Juan basin and the drainage area that provided recharge are larger by several orders of magnitude than the modern flow systems. This allows for discharge of deep, regional flow.

Adding excess head due to dissolved salts reverses the vertical direction of flow in the playa. Under constant density conditions, vertical flux is upward (discharging) in the playa; adding the head due to solutes forces the flux to be downward (recharging). Such variability of vertical flux in playas is verified by modern examples where reversals in flux are associated with climatic changes from wet to dry over a time period of decades (Allison and Barnes, 1985). The model and the natural examples indicate that playa-lake water in the Brushy Basin Member may have moved downward or upward depending on changes in the various factors that affect the density of playa-lake water and subsurface pore water. Such factors would include surface input of fresh water, direct precipitation, dissolution of salt crust, rate of evaporation, and composition of deep gravity-driven ground water.

Deep regional flow and compaction

Varying the upward flow of deep gravity-driven and compaction flow affects the areas of recharge and discharge by increasing both the rate and areal extent of discharge (Fig. 7). The model run illustrated in Figure 7 is identical to that of Figure 5B except for a fivefold increase in discharge from below in Figure 7. In this run, upward discharge from below was set to yield the maximum probable flux, given the overall constraints on the model from precipitation, infiltra-

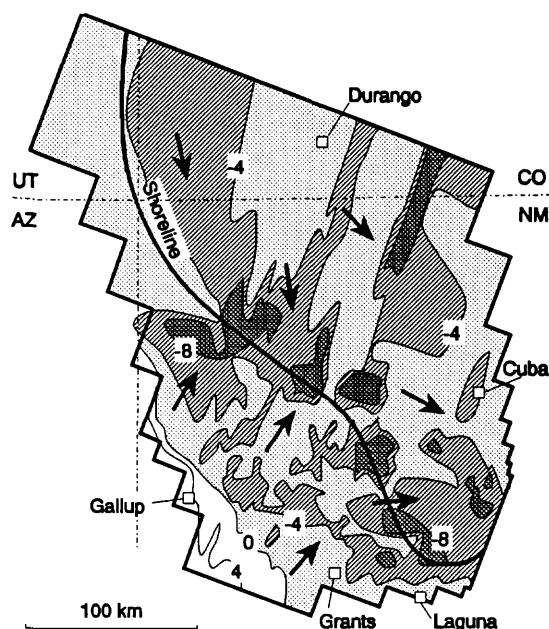


FIG. 7. Map showing effect of increasing flux from deep gravity-driven flow and compaction through base of model. Flux is set equal to maximum probable flux as constrained by climate and recharge. Input is the same as for Figure 5B, except for flux from below, which is increased fivefold.

tion, and recharge as discussed above. Ground water discharges in most of the model area. Maximum discharge is again concentrated along the shoreline. Local hydraulic conductivity variations yield patchy variations in discharge.

The amount of pore water trapped in the sediments places a limit on the amount of ground water available from compaction. Despite the popularity of compaction as a significant mechanism for downward or lateral flow of ground water in the Morrison Formation (Wood, 1968; Shawe, 1976; Adams and Sauer, 1981; Turner-Peterson, 1985; Adams, 1986; Hansley, 1986; Turner-Peterson and Fishman, 1986; Northrop and Goldhaber, 1990; Wanty et al., 1990), simple calculations show that compaction is a minor influence compared to gravity-driven flow (Sanford, 1990a). An estimate of the volume of pore water expelled from a vertical column one meter square can be made using observed porosity-depth curves (e.g., Hanor, 1979; Bond and Kominz, 1984) and stratigraphic thicknesses. For the Brushy Basin Member, having a thickness of 100 m, a porosity of 0.50 at the top, and a porosity of 0.15 at the bottom, the total amount of water expelled is about $100 \text{ m} \times (0.50 - 0.15)/2$ or 17.5 m. (Units are volume of fluid in m^3 expelled per unit area in m^2 .) In order to maximize the contribution from compaction, a minimum reasonable time of 5 m.y. is taken as the duration of deposition of the Brushy Basin Member. This figure comes

from the estimated ages of 149 to 145 ± 1 Ma for about two-thirds of the Brushy Basin section (Kowallis et al., 1991). If the Brushy Basin Member had been deposited in 5 m.y., then the water of compaction contributed an average of 0.0035 mm/yr to the vertical flow. The vertical flow rate for gravity-driven flow of 15 ± 10 mm/yr is therefore approximately 4,300 times greater than that for compaction-driven flow.

For sub-Morrison Paleozoic and Mesozoic rocks, having a thickness of 3,600 m, a porosity of 0.40 at the top, and a porosity of 0.20 at the base, the corresponding amount of pore water expelled is about $3,600 \text{ m} \times (0.40 - 0.20)/2$ or 360 m. This contributes 0.07 mm/yr to the vertical flow, 20 times that from the Brushy Basin. Despite the Brushy Basin muds having greater porosity than the lithified sediments below, the much greater volume of these sediments yielded many times more pore water from compaction. Nevertheless, compaction flow is still minor compared to gravity-driven flow. Although these calculations are greatly simplified, they illustrate the point that there simply was very little pore water available to be compacted out compared to the continuous flow of gravity-driven ground water.

The dominance of gravity-driven flow over compaction-driven flow in this type of basin is also demonstrated by more elaborate calculations. Bethke (1986) estimates maximum compaction-driven flow in the Illinois basin of 2 mm/yr whereas the maximum gravity-driven flow is 11,000 mm/yr, a difference of 5,500 times. Person et al. (1992) estimate maximum compaction-driven flow in the Cooper and Eromanga basins of Australia to be 0.5 mm/yr compared to a maximum gravity-driven flow of 9,289 mm/yr, a difference of over 18,000 times. Thus, available evidence indicates that gravity-driven flow in these settings exceeds compaction-driven flow by some three to four orders of magnitude. In a review of hydrologic literature, Kreitler (1989) concluded that compaction-driven flow may be important in relatively young basins with thick shale sequences; however, gravity-driven flow predominates in mature, topographically uplifted sedimentary basins.

Local tectonics and paleotopography

Local syndepositional tectonic activity affected paleotopography and sedimentation in the Westwater Canyon Member (Kirk and Condon, 1986). Positive structures are interpreted to be elongate domes or folds possibly resulting from reactivation of fault-bounded basement blocks. Sediment distribution over these structures shows isopach thins, low sandstone/mudstone ratios, low net sandstone values, and low percent sandstone in the Westwater Canyon Member and low net sandstone and isopach thins in the overlying Brushy Basin Member. Separating the

highs were low areas or "depocenters" characterized by thick buildups of highly transmissive, clean sandstone.

In order to examine the effect of the local paleotopographic variations described by Kirk and Condon (1986), their actively rising structures were modeled as slightly elevated hills rising 1 to 2 m above the surrounding plain. The hills were located in the area of general ground-water discharge. The calculated effect of hills was to reverse the vertical component of flow and cause local recharge. For a hill (and water table) height of 2 m, the calculated recharge was 13 mm/yr (4×10^{-10} m/s). The locally recharged ground water then moved laterally into the channels in the depocenters, or paleotopographic lows, where it discharged. Thus, the effect of local paleotopographic relief was to set up local flow systems that were superimposed on the regional flow system.

Vertical hydraulic conductivity and transmissivity

Changing the vertical hydraulic conductivity changes the total flow through the model but does not greatly affect the areas of recharge and discharge (Fig. 8). When the hydraulic conductivities of all the cells are multiplied by the same factor, the contours exhibit the same pattern and only the flow rates change. When the difference or differential between hydraulic conductivities of sandstone and mudstone is increased, the patterns of recharge and discharge are similar, but there tends to be more relief in the

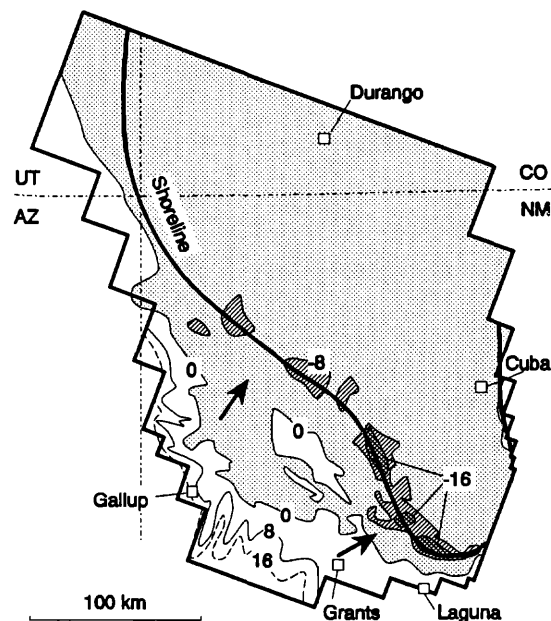


FIG. 8. Map showing effect of increased vertical hydraulic conductivity of confining layers by one order of magnitude compared to Figure 5B.

contour map of recharge and discharge. That is, the local variations in flow rates increase. The model run illustrated in Figure 8 is identical to that of Figure 5B except that the vertical hydraulic conductivity of mudstone layers is increased from 10^{-9} to 10^{-8} m/s. Comparison of the two figures shows that a tenfold increase in the vertical hydraulic conductivity causes approximately a twofold increase in the vertical flux. There are slight changes locally in the areas of recharge and discharge, but the overall pattern is relatively constant. Similar results were obtained over the entire range tested. In general, the total flux through the system and the discharge rate increase with increasing hydraulic conductivity, but the areas of recharge and discharge remain constant. The confining layers apparently are the rate-limiting factor in the total flux, because changing the transmissivity of sandstone aquifers had no significant effect on either flow rates or directions. The transmissivity of the sandstone aquifers only affected the total flux when the permeability of the mudstone units was set to within three orders of magnitude of the sandstone permeability. Because actual permeability contrasts are in the range 10^{-6} to 10^{-12} , it is likely that the confining units controlled the total flow through the system.

To simulate the effect of variable permeability with depth, the Brushy Basin Member was given a higher permeability, by a factor of 10, than the other, deeper confining units. As with the other runs with variable vertical hydraulic conductivity, the areas of recharge and discharge were unchanged, but the rate of flow through the model was increased.

The upper limit on the vertical hydraulic conductivity is set by the available recharge, as discussed above. Model testing showed that 10^{-8} m/s was about the maximum allowable vertical hydraulic conductivity. Greater hydraulic conductivity would lead to ground-water flow that exceeded the amount available from recharge.

Basin margin boundary conditions

The basin margin in the south and southwest was a high area during deposition of the Morrison Formation. The margins were also uplifted after the end of Brushy Basin Member deposition. This has resulted in erosion of the upper parts of the Brushy Basin Member in the south and southwest. The timing of uplift around the basin margins relative to uranium ore formation is unknown. Because uplift in the recharge area could be expected to affect the ground-water flow, the model was run with three different scenarios to simulate the range of possible effects. First, to simulate the time before the Brushy Basin Member was beveled, wherever the thickness of the Brushy Basin Member was less than 30 m, it was set to 30 m. This simulated a reconstructed end-member

condition where the Brushy Basin Member was a maximum. Two other simulations reproduced the opposite condition, where the Westwater Canyon Member and underlying units, including the San Rafael Group, were exposed at the margins of the basin. In one case the exposure was confined to the south, and in the other case the area of exposure extended around both the south and southwest sides of the basin (Fig. 9). The model run illustrated in Figure 9 is identical to that of Figure 5A except that, instead of a no-flow boundary all around the edges of the model, the edges on the south and southwest have constant-head boundaries simulating outcrop of the deeper units.

In all three simulations of different boundary conditions, the effect on the model results was negligible. Neither flow rates nor areas of recharge and discharge were significantly affected. There is some evidence of slightly less discharge in the area just northeast of Gallup, but the areas of discharge are unchanged. This is further evidence that the confining layers control the flow through the system. Thus, uncertainties in the details of uplift and erosion in the highland areas to the south and southwest will not affect the conclusions about the flow of ground water in the basin.

To summarize the controls on ground-water flow, two types of controls were present, those that affected the recharge and discharge areas, and those that did not. Factors primarily affecting the direc-

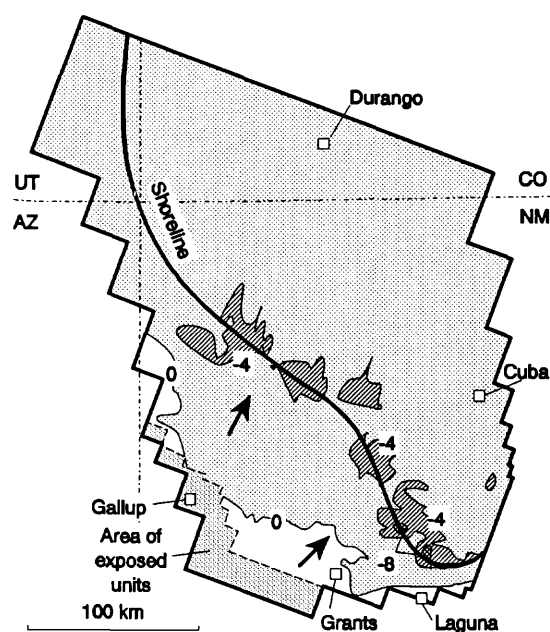


FIG. 9. Map showing effect of changing boundary conditions representing outcrop of units. Input conditions are the same as for Figure 5B, except that units are exposed on the south and southwestern margins of the basin.

tions of flow and recharge-discharge areas are variations in topographic slope, shoreline position, and density contrasts in the lake and pore fluids (excess head due to dense fluid). Input parameters that primarily affect the magnitude of flow are vertical and lateral hydraulic conductivity, topographic slope, depth of erosion in the recharge area, magnitude of deep gravity-driven flow, and rate of compaction. Increasing the value of any of the second set of parameters increases the throughput of the system. The location of discharge and recharge areas, however, is relatively insensitive to variations in these parameters.

Favored models for flow during uranium deposition

Reconstructed ground-water flow in the Westwater Canyon and Jackpile Members conforms to the above hydrologic constraints (Fig. 10) and is within the range of ages proposed for uranium deposition. For the model run simulating ground-water flow in the Westwater Canyon Member (Fig. 10A), the hydraulic conductivity of sandstone and mudstone is 10^{-3} and 10^{-8} m/s, respectively; the lake shoreline is at the transition between carbonate mud flat and clinoptilolite facies; the density of the alkaline-saline lake water increases the equivalent fresh-water potential by 6 m above the lake level in the center of the basin; Middle and Upper Jurassic sedimentary rocks are exposed at the surface on the southern margin of the basin; the Brushy Basin Member was deposited

over a period of 10 m.y.; and deep gravity-driven flux was adjusted to yield an average basinwide discharge rate of 10 mm/yr. This discharge rate was calculated from the 10:1 ratio of recharge to discharge area combined with empirical estimates of recharge rates in similar modern environments, as discussed above.

The model cannot determine the age of ore formation, but it can indicate the flow, assuming different ages. Figure 10A illustrates the hydrologic setting assuming that the uranium deposits in the Westwater Canyon Member formed either during deposition of the Brushy Basin Member when the shoreline of the playa lake was in the mud flat zone or during deposition of the Westwater Canyon Member where the shoreline marks an abrupt change in topographic slope. In this setting, upward discharge would have been highly focused during ore formation.

The results of the present model can also be applied to ore deposit formation at different times. If ore formation had occurred earlier, during deposition of the host Westwater Canyon Member, and there was no sharp change in topographic slope but only a gradual transition, then discharge would have occurred over a broader zone where alluvial plain merged into mud flat, and ore deposits would have formed in a broader zone of ground-water discharge. If ore formation had occurred after deposition of the Brushy Basin Member, then they formed near the hinge line, in a zone of largely horizontal flow (com-

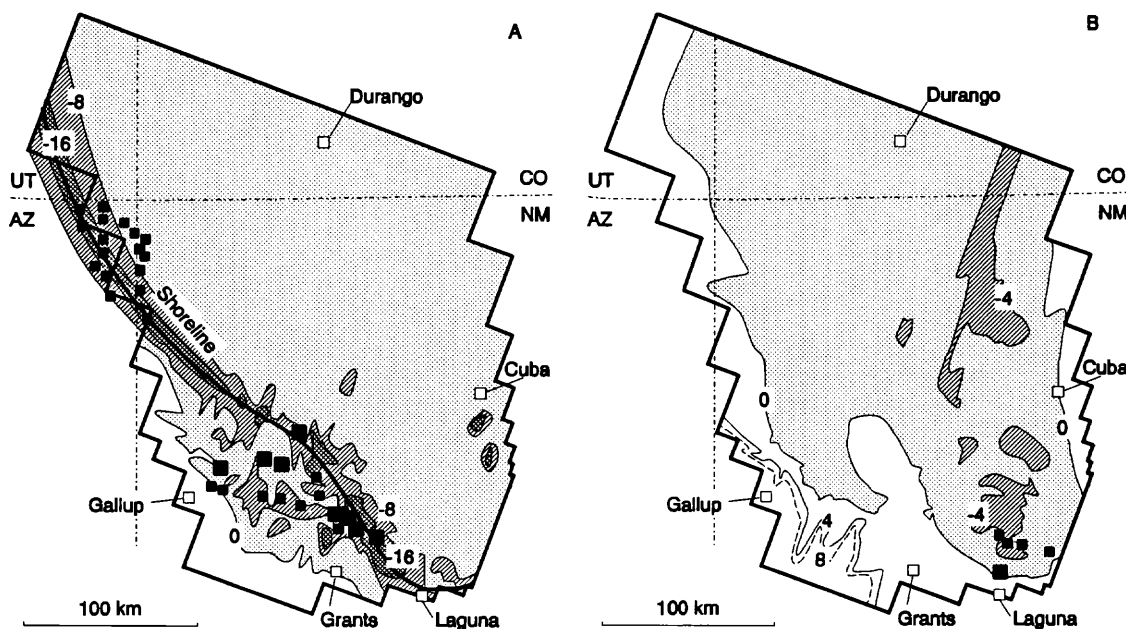


FIG. 10. Maps showing inferred ground-water flow during uranium ore deposition in A, the Westwater Canyon (and Salt Wash Member in northwestern part of San Juan basin), and B, the Jackpile Members. Large squares indicate uranium deposit clusters having more than 10,000 short tons of U_3O_8 , small squares, less than 10,000 short tons (Finch, 1991). Note that the main areas of ground-water discharge coincide with the location of uranium deposits.

pare location of uranium deposits in Fig. 10A with flow in Figs. 5A or 10B). In all cases, the uranium deposits formed in a area of discharge.

On the basis of the present model, simulated recharge to the Westwater Canyon Member was from three main sources. Shallow recharge moved down-dip to the northeast. At the lake shoreline, it rose and discharged. Deep gravity- and (minor) compaction-driven flow moved up through the base of the Morrison Formation, converged, and discharged at the lake margin. Because of the underlying evaporites, this deep fluid was probably saline, as suggested by modern analyses (Sanford, 1990a). Despite the tendency for alkaline-saline playa-lake water to move downward, upward-moving deep gravity-driven flow was usually more significant and the net effect was upward discharge through the lake bed. In some simulations, density-driven playa-lake water descended through the lake bed, moved outward through the Westwater Canyon, then rose and discharged at the playa-lake margin. This recirculating playa-lake water would have come in contact with deep, regional saline ground water rather than dilute local ground water. Calculated horizontal fluxes in the Westwater Canyon Member were about 1 to 5 m/yr for northeastward fresh-water flow and about 0.1 to 1 m/yr for outward playa-lake salt-water flow. The vertical discharge rate was about 10 mm/yr. As discussed above, these ground-water flow rates were set to resemble those in analogous modern lacustrine and marine environments (e.g., Duffy and Al-Hassan, 1988; Sanford and Konikow, 1989).

Simulated ground-water flow during and shortly after deposition of the Jackpile Member (Fig. 10B) was calculated for the conditions shown in Figure 5B. Simulated recharge is from two sources. Shallow gravity-driven ground water recharged the Jackpile Member from the south and southwest. Deep gravity-driven flow moved up through the Brushy Basin Member. The deeper ground water was probably saline or alkaline-saline and formed a subhorizontal interface with the overlying dilute ground water. Early discharge probably would have a large proportion of alkaline-saline pore water owing to devitrification and alteration of tuffaceous Brushy Basin lake sediments; later discharge probably would have a larger proportion of saline ground water derived from deep interaction with evaporitic rocks beneath the Morrison Formation. Ground water discharged throughout the playa with most of the discharge at the lowest elevations. Horizontal flow rates were about 1 to 5 m/yr in the Westwater Canyon Member and Jackpile Member, and vertical discharge was about 10 mm/yr, based on modern analogues.

Conclusions

The results show that tabular sandstone uranium deposits formed in areas of regional ground-water

discharge (Fig. 10). In the Westwater Canyon Member and Grants uranium region, discharge was localized along the shoreline of the Brushy Basin lake when present. If no lake was present, discharge occurred where the topographic slope flattened significantly where the alluvial plain merged into mud flat and at the lowest elevation in the basin. Where syn-depositional tectonic activity caused local paleotopographic highs and lows, local ground water recharged at the highs and discharged at the lows, even reversing the regional flow pattern locally. As shown by Kirk and Condon (1986), the paleotopographic lows or depocenters were favorable for uranium deposits. Thus, the uranium deposits formed in the lows where both regional and local ground water discharged. In the Jackpile Member, discharge was concentrated in the lowest parts of the basin where the Jackpile Member was deposited in an active syncline. In both cases, uranium deposits formed in areas of major regional and local ground-water discharge.

The spatial correlation between tabular sandstone uranium deposits and regional ground-water discharge in the San Juan basin is robust. All likely parameters controlling regional ground-water flow were tested in the course of the modeling. In most cases, the basin from the distal edge of the alluvial plain to the center was the site of discharge, and the shoreline area was the site of maximum discharge. At times, the basin center was an area of recharge for alkaline-saline playa-lake water. There were no conditions where ground water recharged at the basin margin. These computer results are confirmed by modern analogues where playa margins are typically zones of ground-water discharge (e.g., Allison and Barnes, 1985; Mono Basin Ecosystem Study Committee, 1987; Duffy and Al-Hassan, 1988; Macumber, 1992).

These results place strict constraints on the source of uranium and associated humate. Uranium and humate could not have been derived from a source down the hydraulic gradient from the ore deposit, such as the Brushy Basin Member at the shoreline. Uranium may have come from updip in the shallow aquifer or it may have come from deep regional flow. Humate is not very soluble in brine, so it probably was not carried by the deep regional flow. Its source therefore appears to be the host sediments laterally adjacent to or updip from the ore deposits. Local, transient, fresh-water recharge may have been important for transport of dissolved organic matter and should be incorporated into future models.

According to the present model, shallow ground water moved laterally down-dip and upward in the sandstone units. The subhorizontal orientation and tabular shape suggest a fluid interface. Many uranium-humate and uranium-vanadium deposits are elongated along the channels and rise stratigraphically toward the basin (Hazlett and Kreek, 1963;

Reinhardt, 1963; Moench and Schlee, 1967; Granger, 1968; Hilpert, 1969; Fischer, 1974; Melvin, 1976; Northrop and Goldhaber, 1990). Both features suggest moving ground water and are consistent with an interface. Stratigraphic rise is expected at an interface in a zone of ground-water discharge (Hubbert, 1940). The present modeling suggests that this interface existed where shallow, fresh water rested upon deep, saline ground water. The saline ground water recharged as fresh water far to the south and west and interacted with evaporitic rocks along the flow path; the saline ground water was not the alkaline-saline pore water from the Brushy Basin lake.

Little drilling for uranium has taken place in the deeper part of the section toward the center of the basin, so the relationship between uranium and ground water is unknown. The present results suggest that uranium would be found further into the basin if the appropriate fluvial sandstone were present.

A complete explanation for the association between ground-water discharge and uranium deposition is beyond the scope of this paper, which only addresses the question of the direction of ground-water flow during uranium deposition. However, the present model has important implications for previously proposed models that are primarily based on field relations and diagenetic evidence.

Fischer (1937) concluded that the concentration of uranium in tabular deposits was virtually syngenetic and proposed that bacterial degradation of organic matter caused reduction of uranium and other metals. Later, he suggested that downdip flow precipitated uranium at a fluid interface (Fischer, 1947). Many early workers favored gravity-driven flow rather than compaction because of the larger possible flow rates (Fischer, 1947; Hilpert, 1969). All these suggestions appear to be essentially correct and are completely consistent with the hydrology proposed here. Instead of a water table, which would imply oxidized rock above the interface, the interface was most likely between fresh and saline ground water, which would be compatible with reduced rock above and below. Bacterial action is highly likely, particularly since the discovery that dissolved organic matter can precipitate as humate at a saline water interface (Granger et al., 1961; Swanson and Palacas, 1965; Ortiz et al., 1980). A syngenetic source of humate and at least some of the uranium is compatible with the present model.

Hypotheses calling for late epigenetic concentration of organic matter and/or uranium by downward-moving solutions from the Upper Cretaceous Mancos Shale (Shawe, 1956, 1976) or Early Cretaceous swamps that predated the Upper Cretaceous Dakota Sandstone (Granger et al., 1961) are unlikely, at least as an essential process in the formation of tabular

sandstone uranium deposits in general. As discussed above, compaction is volumetrically negligible compared to gravity-driven flow, and it is generally upward and outward. The downward and downdip flow required by these models is hydrologically difficult to explain. Although Shawe (1976) calls upon faults to increase fluid flow, and Granger (1968) calls upon erosion that exposed the Westwater Canyon Member to recharge, downward flow through the clay-rich Brushy Basin Member is particularly unlikely. There is no evidence relating sedimentary facies in either the Mancos or Dakota to uranium deposits. The probable high salinity of fluids expelled from the marine sediments would not favor transport of dissolved organic matter from the Mancos. A Cretaceous age is inconsistent with evidence for nearly syngenetic humate and/or uranium precipitation (Squyres, 1980). Although both Shawe (1956, 1976) and Granger et al. (1961) advocate a fluid interface, only a short-lived, essentially syngenetic, fluid interface is hydrologically very likely. Evidence suggesting mineralization long after sedimentation (Shawe, 1956, 1976) may be due to reduction at preexisting tabular concentrations of organic matter that have since been destroyed or to remobilization associated with faults. The deposits in the Slick Rock district described by Shawe (1976) may also have had a locally more complex diagenetic history than those in the San Juan basin, as suggested by Hansley and Spirakis (1992).

The results of this paper also cast doubt on significant elements of the "lacustrine-humate" model of Turner-Peterson and others (Peterson, 1980; Turner-Peterson, 1985; Adams, 1986; Turner-Fishman and Turner-Peterson, 1986; Peterson and Fishman, 1986;). According to the lacustrine-humate model, ground water was squeezed by compaction out of mudstones in the mud flat of the Brushy Basin lake and moved downward into underlying sandstones to form the uranium deposits and cause extensive alteration and dissolution of the host rocks (except in the Jackpile Member, where water of compaction was squeezed out and upward). In fact, the mud flat facies of the Brushy Basin Member was the least likely place for downward flow of ground water during sedimentation. Hydrologically, there are three major problems with the lacustrine-humate model. First, water of compaction is a very minor constituent of ground-water flow compared to gravity-driven flow. As discussed above and elsewhere (Bethke, 1986; Sanford, 1990a), compaction typically is two to four orders of magnitude less than gravity-driven flow under these conditions. There was probably not enough water of compaction to account for the extensive alteration that reaches to the base of the underlying Westwater Canyon Member. Second, the margins of lakes, especially the margins on the upstream side of large playa lakes are zones of ground-water discharge (Habermehl, 1980; Allison

and Barnes, 1985; Pfannkuch and Winter, 1985; Mono Basin Ecosystem Study Committee, 1987; Duffy and Al-Hassan, 1988, Winter, 1988; Winter and Woo, 1990; Macumber, 1992). Based on extensive observation of modern systems and on modeling, it is virtually inconceivable that the shoreline of the Brushy Basin Member would have been dominated by downward- and outward-flowing pore water. The minor quantities of expelled, compaction-driven pore water would be vastly overwhelmed by discharging, gravity-driven ground water. If the compaction assumed by the lacustrine-humate model took place after deposition of the Brushy Basin Member, pore water expulsion would have been even less significant. During the post-Brushy Basin-pre-Dakota period of minor fluvial deposition, non-deposition, and erosion, gravity-driven flow would have been as great or greater than during deposition of the Brushy Basin Member. Meanwhile, sedimentation was slow to nonexistent, so that compaction would have diminished. Plus, most pore water in mudstones is normally lost during early burial. Thus, after deposition of the Brushy Basin Member, compaction would have decreased whereas gravity-driven flow would have remained the same or increased. Third, not only would downdip, gravity-driven flow discharge at the playa-lake margin, but recirculating, outward-flowing playa-lake water would also discharge at the lake margin (Figs. 6 and 10A). Turner-Peterson and Fishman (1986) recognize that playa-lake water may be driven downward by density into the underlying sediments. Diagenetic alteration in parts of the Westwater Canyon Member is consistent with such downward flow from the central playa of the Brushy Basin Member (Hansley, 1986, 1990). Turner-Peterson and Fishman (1986) are unclear where this recharging playa-lake water ultimately would go. Hydrologic modeling of modern environments indicates that the playa-lake water moves outward toward the playa margins where it mixes with fresher ground water and discharges at the shoreline or playa margin (Duffy and Al-Hassan, 1988). After deposition of the Brushy Basin Member and disappearance of the playa lake, ground-water flow would have been gravity-driven and downdip. Discharge would have dominated in all the areas at low elevations (Figs. 5A and 10B). Arguments by authors of the lacustrine-humate model against the presence of a brine (Turner-Peterson, 1985; Fishman and Turner-Peterson, 1986) rest on unproven assumptions that the humate originated in the Brushy Basin Member and that albitization was caused by the same brine that formed the fluid interface. Without these assumptions, there is no valid argument against a brine being present and considerable evidence for its presence.

The present results are largely consistent with the model of Northrop (1982), Northrop and Goldhaber

(1990), and Wanty et al. (1990) for uranium deposits in the Salt Wash Member of the Henry basin of southwest Utah but provide additional constraints on the timing of ore formation. These authors interpreted petrographic, chemical, and isotopic evidence to indicate a brine-fresh water interface where bacterial degradation of detrital, cellular, organic matter and reduction of sulfate from the underlying brine generated reduced sulfur species that precipitated uranium and vanadium. Mixing reactions at the interface caused precipitation of dolomite and quartz overgrowths associated with and extending beyond uranium ore layers. The hydrologic modeling presented here, as well as observations of modern hydrologic systems, are compatible with a brine interface; however, the sensitivity of hydrologic systems to subtle tectonic, paleotopographic, and climatic changes suggests that an interface at any one location would be short-lived. Given the typical rates of such changes, an interface may have lasted for 10^3 to 10^5 yr but probably not more than 10^6 yr. Based on dating of uranium ore, Wanty et al. (1990) suggest that an interface was present at 115 Ma, about 35 m.y. after sedimentation of the host sandstone. During this 35-m.y. time period, the area was covered by the Brushy Basin Member, slightly uplifted, and then covered by the fluvial Cedar Mountain Formation (Barremian(?) to Albian; Tschudy et al., 1984; Molenaar and Cobban, 1991; W.M. Aubrey, U.S. Geological Survey, oral commun., 1993). Major changes in hydrologic flow during this time would have made a long-lived interface unlikely. A short-lived interface at 115 Ma is possible, but then the relationship between uranium deposits and sedimentary facies in the host sandstone (Peterson, 1980; Peterson and Turner-Peterson, 1980; Sanford, in prep.) is unexplained and recharge of dilute ground water down through the Brushy Basin Member is a problem hydrologically. Further, Spirakis (1991) and Hansley and Spirakis (1992) point out that formation of quartz overgrowths and vanadium as proposed by Wanty et al. (1990) and Goldhaber et al. (1991) require unreasonably long times or high flow rates. An early, short-lived interface, complexing of silica by organic acids, and accumulation of uranium and vanadium on or near preexisting organic matter over a long time would help to resolve these difficulties.

The diagenetic processes inferred by Hansley and Spirakis (1992) are consistent with the hydrogeology presented here, but some of their conclusions have not been fully verified. According to their model, shallow, relatively fresh, mildly alkaline ground water carries humic acid downward and meets a deeper, saline ground water at a density-stratified interface where increased salinity causes precipitation of humic acid as humate by cation loading in subhorizontal tabular layers. During and after humate deposition, uranium precipitates by reaction with both dis-

solved humic acid and precipitated humate. With deeper burial, the uranium-humate ore layer is then diagenetically modified by fermentation (methanogenesis) and organic acid production. Organic acids are produced at high temperatures ($>80^{\circ}\text{C}$) by thermal degradation in the absence of bacterial action. Among other diagenetic reactions, humate is converted into a mass of coffinite, clay minerals, and minor organic residue, and dolomite and silica precipitate. Unlike organic matter in the Grants mineral region, all of the organic matter studied in the Henry basin is cellular and not amorphous (Northrop and Goldhaber, 1990). Hansley and Spirakis (1992) present evidence that diagenetic reactions consumed most of the original amorphous organic matter in the Henry basin and Slick Rock areas. Hansley and Spirakis (1992) argue that dolomite formed in part from organic carbon. In order to yield the observed carbon isotope ratios, such a process would require a fortuitous combination of conditions that is much less likely than the straightforward remobilization of carbonate from underlying evaporites as suggested by Northrop and Goldhaber (1990). The nearly complete degradation of organic matter suggested by Hansley and Spirakis (1992) is plausible, but evidence for incorporation of organic carbon in the dolomite is inconclusive.

The present model recreates ground-water flow during deposition of the Brushy Basin Member (and the Westwater Canyon Member), but the results do not indicate that the uranium deposits actually formed during Brushy Basin deposition. Extensive lithofacies and petrographic evidence suggests that the location of uranium deposits was controlled by hydrologic conditions during sedimentation of the host sandstone (Sanford, in prep.). In fact, lake-related facies of the Brushy Basin Member probably did not control the location of Colorado Plateau uranium deposits, because the correlation between uranium deposits and mud flat facies of the Brushy Basin Member (Turner-Peterson, 1985; Turner-Peterson and Fishman, 1986) does not hold for the Colorado Plateau outside of the Grants uranium region (Sanford, in prep.).

It is quite possible that humate layers precipitated very early and that uranium accumulated much later (Levanthal and Granger, 1977; Squyres, 1980). Tabular humate layers may have formed contemporaneously with sedimentation of the Westwater Canyon Member in a brief period of time during which a fluid interface was present. During a much longer period of time, perhaps millions of years, uranium may have accumulated on the existing humate and detrital organic matter. An interface may not have been present when the uranium deposits formed, and the tabular configuration of the uranium may be due to preexisting conditions.

The use of paleohydrologic analysis can be a key to

unlocking the mystery of the formation and localization of tabular sandstone uranium deposits. Although the deposits had been studied for more than four decades by geologists using standard stratigraphic, sedimentologic, and petrologic tools, none of these methods yielded a clear answer to the question of the patterns of ground-water flow during ore deposition. By applying the techniques of modern ground-water analysis, the pattern of regional ground-water flow can be better understood. Many details remain unsolved, some of which can be addressed by models that incorporate transient and variable-density flow. In light of this progress, modern methods of paleohydrologic analysis should be considered as a standard tool in the analysis of a wide variety of ore deposit types as well as in the exploration for new deposits.

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