GROUNDWATER VELOCITIES AT THE NEVADA TEST SITE: ^{14}C CARBON-BASED ESTIMATES

Prepared by

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INTRODUCTION

Chemical and isotopic data can be used to constrain and validate groundwater flow models. This study examines probable groundwater flowpaths at the Nevada Test Site (NTS) and estimates groundwater velocities for these flowpaths using water chemistry and carbon isotopes. These velocities are provided for comparison to velocities calculated by a numerical flow model developed by GeoTrans, Inc. Similar to numerical flow models, models of chemical and isotopic evolution are not unique; any number of combinations of reactions can simulate evolution from one water to another, but are no guarantee that the simulation is correct. Knowledge of the hydrology, mineralogy, and chemistry must be combined to produce feasible evolutionary paths.

There were six steps followed in this study, each building upon the previous ones: 1) evaluating the locations of carbon isotope groundwater analyses, including availability of a reliable ionic analysis and well completion information; 2) pairing wells that could be reasonably expected to lie along a flowpath, considering gross flow directions, hydraulic barriers, and producing horizons; 3) judging whether a flowpath was isolated or experiencing mixing of waters from different sources; 4) running chemical speciation models to determine the saturation state of mineral phases, and gathering information on aquifer mineralogy; 5) simulating the carbon chemical evolution along the flowpath with NETPATH (Plummer *et al.*, 1991), using the measured and calculated ^{13}C values as an indicator of fit; and 6) using the ^{14}C data to calculate velocity.

This process is described below. First, the wells and flowpaths considered are described and the rationale presented for the ones selected for analysis. This is followed by separate descriptions of the chemical and isotopic evolution for the three flowpaths selected (Figure 1): 1) in the volcanic aquifer in Fortymile Wash, 2) the carbonate aquifer of northern Yucca Flat, and 3) the carbonate aquifer of southern Yucca Flat. The report concludes with a discussion of the uncertainty in the velocity estimates and suggestions for reducing that uncertainty.

SPATIAL DISTRIBUTION OF ^{14}C CARBON DATA AND THEIR RELATIONSHIP TO POSSIBLE FLOW SYSTEMS

Fortymile Wash Flowpath

Several wells are positioned along Fortymile Wash and have carbon data: Ue29a#2, J-13, J-12, and Lathrop Wells (15S/50E-18cdc). Ue29a#2 and the J wells are completed in tuff. The geology for Lathrop Wells has not been found, but is likely to be tuffaceous alluvium. Multiple, and similar, analyses are available from these wells, lending confidence in the data. The flowpath investigated is Ue29a#2→J-13/J-12→Lathrop Wells. The flowpath from J-13 to J-12 cannot be modeled because J-12 has a higher amount of ^{14}C , despite being downgradient. This is likely to be a result of the three-dimensional nature of the flow system, with J-12 being shallower and accessing younger water than present deeper in the flow system.

Carbonate Flowpaths

The regional carbonate aquifer underlying a vast area of southern Nevada is often casually considered to be a continuous hydrogeologic unit and, when considering large-scale interbasin flow, such a representation may be adequate. However, on smaller scales and when considering transport in addition to flow, the carbonate system is much more complex, with compartmentalized flow. The

Figure 1. Location map showing the wells with ^{14}C data evaluated for this study and the possible flowpaths selected for analysis. Only wells completed in aquifers along Fortymile Wash, on Pahute Mesa, and in the regional carbonate aquifer were considered. Many additional wells have ^{14}C data at and near the NTS but were not within the scope of this work.

most thorough investigation of the nature of the regional carbonate aquifer in the vicinity of the NTS was performed by Winograd and Thordarson (1975). These researchers delineated the Lower Carbonate Aquifer in the NTS area and demonstrated the existence of hydraulic barriers within the carbonates, suggesting that such barriers are numerous, compartmentalizing the aquifer into regions bounded by faults or clastic rock barriers. Adding to the uncertainty is the relationship with

overlying units, as Winograd and Thordarson (1975) suggested that downward vertical leakage into the Lower Carbonate Aquifer may be enhanced along the margin of the basin and in the vicinity of faults. Compounding the hydrogeologic complexity are variable, and sometimes unknown, well completions. These factors combine in the NTS data such that wells that are apparently along reasonable carbonate flowpaths may in fact not be closely hydrologically related to one another.

Eleven wells have been identified for possible chemical and isotopic analysis of carbonate flowpath characteristics (Table 1). Each of these is discussed below.

Test Well 2 (TW2). This well is completed in the Pogonip Group and produces from fractured dolomite at an elevation of 322.5 to 398.1 m relative to mean sea level. The hydraulic head elevation was originally about 736 m, but has been depressed more than a hundred meters by pumping. One concern with this well is that it was acidized to increase production, but the ^{13}C values do not seem unusually heavy as would be expected if dissolved carbonate rock had contributed significantly to the dissolved carbon. There is some difference between the two carbon analyses available from this well; it may be that the difference reflects different water discharged by the well through time, as the samples are many years apart (1969 and 1982) and this well has been pumped extensively for water supply.

Test Well D (TWD). This well has a number of factors that raise concern regarding interpreting samples. The well is reported as being in Paleozoic limestone, dolomite, and argillite, with a notation that it might be completed in the "Eleana (?)" Formation. Indeed, the well is located immediately at the eastern edge of the Upper Clastic Aquitard region delineated by Winograd and Thordarson (1975). In this location, the well may receive little flow from the north, such as the area around TW2; instead, water may originate from vertical leakage or from areas of much higher hydraulic head to the west. Of even more concern is the well completion open to the overlying volcanic units, as well as whatever carbonate is below.

Ue1b and Ue1c. These wells access either the Eleana Formation or Upper Carbonate Aquifer Tippipah Limestone and thus should not be along direct flowpaths with Lower Carbonate Aquifer wells. A flowpath from Ue1c to Test Well D may be reasonable to evaluate eastward flow into Yucca Flat, but not strictly within the Lower Carbonate Aquifer itself.

Ue1q. This well is now open to Paleozoic carbonates because of recent recompletion work. The elevation of producing carbonates is 451.7 to 494.7 m, with a head elevation of 739 m. Though this reported head is higher than at TW2, the true head relationship between these wells is uncertain. The surface elevation at Ue1q may have changed due to subsidence caused by underground testing. It is assumed here that flow can occur from TW2 to Ue1q, consistent with Winograd and Thordarson's (1975) general flow directions. However, it is possible that flow to Ue1q may be from a more westerly direction, similar to that described above for Test Well D. Ue1q is west of Yucca Fault, interpreted by many as a major hydraulic barrier preventing east-west flow in the carbonates. If true, there is not a direct flowpath from Ue1q to ER-6-1. A concern is the great difference between the ^{13}C values reported by Lawrence Livermore National Laboratory (LLNL) and Desert Research Institute (DRI) (Table 1). Although these differences are presently unexplainable, review of historical NTS and regional carbonate aquifer carbon isotope data suggest more consistency, and therefore greater confidence, in the DRI-generated values at this time.

TABLE 1. CARBON ISOTOPE DATA FROM SELECTED WELLS AT THE NTS.

WELL	SAMPLE DATE	SOURCE	$\delta^{13}\text{C}$, ‰	PMC	PMC _{corrected} [†]	AGE _{corrected} , yrs
TW2	1982	Boughton (1986)	-11.2	10	10.71	18,465
TW2	1969	Grove <i>et al.</i> (1969)	-14.52	15	13.14	16,778
TWD	1993	LLNL*	-5.5	2.8	6.11	23,110
Uelb	1992	Davisson <i>et al.</i> (1994)	-4.5	16.0	42.7	7,041
Uelc-1350	1992	Davisson <i>et al.</i> (1994)	-4.6	2.6	6.8	22,245
Uelc-1636	1992	Davisson <i>et al.</i> (1994)	-5.3	2.6	5.9	23,416
Uelq	1992	Davisson <i>et al.</i> (1994)	-2.4	7.7	38.5	7,891
Uelq	1992	DRI**	-8.5	3.8	5.36	24,184
Uelh-1979	1993	LLNL*	-12.1	12.2	12.1	17,460
Uelh-2137	1993	LLNL*	-10.3	23.9	27.84	10,570
ER-6-1	1992	Davisson <i>et al.</i> (1994)	-0.7	2.1	36.0	8,446
ER-6-1	1992	DRI**	-6.3	2.4	4.57	25,507
C-1	1983	Boughton (1986)	-3.8	0.8	2.53	30,409
C	1993	LLNL*	-4.2	0.6	1.71	33,615
Army#1		Spencer (1990)	-5.6	3.1	6.64	22,417
Army#1	1993	Davisson <i>et al.</i> (1994)	-7.1	5.4	9.13	19,791
Am.Tr.2	1973	Spencer (1990)	-4.6	1.8	4.70	25,285
Am.Tr.2		USGS Denver lab rpt***	-7.4	7.3	11.84	17,641
Uel19h	1992	Davisson <i>et al.</i> (1994)	-3.3	9.4	34.18	8,874
Uel19c	1982	Boughton (1986)	-11.6	7.2	7.45	21,471
Uel19c	1992	Davisson <i>et al.</i> (1994)	-5.3	11.4	25.81	11,196
Ue20bh-1	1993	LLNL*	-9.2	21	27.39	10,705
U20a#2		Waddell <i>et al.</i> (1984)		16		
U20a#2	1964	Grove <i>et al.</i> (1969)	-13.5	15.3	13.6	16,493
Uel18r	1992	Davisson <i>et al.</i> (1994)	-1.4	8.2	70.29	2,915
Ue29a#2	1982	Benson <i>et al.</i> (1985)	-13.1	60	54.96	4,948
Ue29a#2	1982	Benson <i>et al.</i> (1985)	-13	62.3	57.51	4,574
J-13	1982	Boughton (1986)	-7.8	29.3	45.08	6,587
J-13	1971	Benson <i>et al.</i> (1985)	-7.3	29.2	48.0	6,068
J-13	1993	LLNL*		30.4		
J-12	1971	Benson <i>et al.</i> (1985)	-7.9	32.2	48.91	5,912
J-12	1993	LLNL*		32		
Lathrop Wells		Spencer (1990)	-7.4	11.7	18.97	13,741

*LLNL is from Lawrence Livermore National Laboratory files

**DRI is from Desert Research Institute files

***rpt is as reported in the GeoTrans carbon database

$$\dagger \text{corrected using the equation } \alpha_{\text{corrected}}^{14} = \frac{\alpha_{\text{measured}}^{14}(\delta^{13}\text{C}_{\text{plants}})}{(\delta^{13}\text{C}_{\text{sample}})} \quad \delta^{13}\text{C}_{\text{plants}} = -12\text{‰}$$

Ue1h. This well is reportedly open to dolomite at an elevation of 194.2 to 699.5 m, with a water elevation of 743.1 m. Though this head is reasonable for the Lower Carbonate Aquifer, the completion of the well and the two (very different) analyses raise concerns. According to well completion records, after casing was placed in the hole, the well was cemented by perforating and squeezing cement into the annulus. Seals emplaced in this manner are often very poor in quality, and it is possible that water from Tertiary units is entering Ue1h. This completion may explain the radically different PMC values of 12 and 28 from samples collected at depths of 603.2 and 651.4 m. Some validation should be performed to evaluate which, if either, water sample is representative of the carbonate groundwater in this well. As with Ue1q, Ue1h is located west of Yucca Fault.

ER-6-1. This well is completed in the Simonson (?) Dolomite at an elevation of 378.0 to 646.5 m. The hydraulic head is 730.0 m. This well is east of Yucca Fault and thus has at least that major structural feature between it and the wells to the northwest. There are no known structures between ER-6-1 and the C wells. As with Ue1q, the ^{13}C values reported by LLNL and DRI are very different. The LLNL value (-0.7) is unusually heavy and will not be used for evaluation.

Wells C and C-1. These wells are completed in dolomitic limestone ranging from elevations of 674.8 to 727.0 m. Hydraulic head is approximately 725.4 m. These supply wells were the site of a tritium tracer test, but no problems should be presented for carbon dating. Carbon data from each well are in good agreement. It has been suggested that the C wells may rest in a carbonate block separated from much of the northern system by thrust faulting at the south end of Yucca Valley, but there are no prominent structural features between ER-6-1 and these wells.

Army #1. This well is completed in limestone and dolomite with minor shale, at elevations of 641.3 to 717.5 m and 366.1 to 546.8 m. Water level elevation is 721.5 m. There is a large distance from this well to other carbonate wells, making the possibility of a direct flowpath solely from the area of the C wells unlikely. The younger age of Army #1 as compared to the C wells requires mixing with a younger water, likely from the HTH-3 area and/or the north end of the Spring Mountains.

Amargosa Tracer #2. This well was completed in carbonates as part of a tracer experiment. There is a large difference between the two reported analyses that would result in a significant difference in groundwater velocities from Army #1 to Amargosa #2. The position of Amargosa #2 closer to the regional discharge area also introduces the possibility of mixing with water from other areas, most notably the northern Spring Mountains, which would affect the groundwater velocity calculations.

Given the inferred major influence of Yucca Fault, flowpaths are selected that do not cross this feature. This leaves a western flowline from TW2→Test Well D→Ue1q→Ue1h. For reasons outlined previously, samples from Test Well D and Ue1h may not represent water flowing from the TW2 area. This leaves TW2→Ue1q. East of Yucca Fault, the best flowpath is from ER-6-1→C wells. Modeling from the C wells to Army #1 and Army #1 to Amargosa Tracer #2 is not recommended at this time because the assumptions regarding mixing with younger water from the northern Spring Mountains would render any calculated velocities non-unique.

Pahute Mesa Flowpath

Carbon-14 data are much more scarce for wells in the Pahute Mesa flow region. Data are available from five wells: Ue19h, Ue19c, Ue20bh-1, U20a#2, and Ue18r. Well completion

information is scanty from Ue19h; records indicate that the well is cased to a depth of 707.4 m (with no record of perforations) and plugged with cement from a depth of 698.9 to 1046.0 m. This would seem to preclude water entry into the well; however, LLNL collected a sample in 1992. With the sampled aquifer unknown, this sample will not be considered further. Ue20bh-1 was never developed after drilling; this history combined with the relatively high PMC value makes the validity of the sample suspect in terms of representing *in situ* groundwater.

Ue19c and U20a#2 are located along a reasonable flowpath, as depicted by Blankennagel and Weir (1973), for generally southwesterly flow across Pahute Mesa, but PMC values are higher at U20a#2 than at Ue19c, suggesting mixing along the flowpath. Ue18r is south of projected flow from U20a#2, but could reasonably be downgradient from Ue19c. The volcanic hydrostratigraphy is too complex to determine if the water from these wells is from the same units. Geologically, Ue19c is completed in the Belted Range Rhyolite, U20a#2 is in Upper Rhyolite Lavas of Area 20, and Ue18r is a very deep well (1525.2 m) completed in tuffs and rhyolite. Neither of the isotopically heavy ^{13}C isotope values reported by LLNL for Ue19c and Ue18r appears reasonable given the predominant volcanic hydrostratigraphy, and the Ue19c value is very different from an earlier DRI value. The LLNL ^{13}C values will not be used because no large source of isotopically heavy carbon is suggested by the well's geologic setting. Given the flowpath depicted by Blankennagel and Weir, the recommended flowpath proposed for analysis on Pahute Mesa would be Ue19c→Ue18r. However, without a valid ^{13}C value for Ue18r, a check on possible chemical and isotopic evolution models is not possible, so no calculations are performed. Preliminary attempts were made in this study to model the chemical and carbon isotopic evolution between Pahute Mesa, well PM-3 (west of the NTS boundary) and Oasis Valley, but no simulations successfully modeled the carbon isotope evolution, presumably because of mixing along these very long flowpaths, and thus the results are not reported below.

RESULTS AND DISCUSSION

Fortymile Wash Flowpath

One of the few areas on the NTS where there is a consensus that significant infiltration of precipitation occurs is along the Fortymile Wash drainage. This flowpath has been the focus of careful geochemical and isotopic research, with the most pertinent to the present question being Claassen (1985) and White and Chuma (1987). These researchers reach different conclusions regarding the flowpath considered here. Claassen concludes that recharge to alluvium in the west-central Amargosa desert occurs primarily by overland flow, so that groundwater sampled at Lathrop Wells and vicinity would be unlikely to have its origin upgradient near the J-wells. White and Chuma interpret essentially the same data differently, concluding that groundwater in upper Amargosa Desert alluvium originated principally as groundwater in tuffs in Fortymile Canyon.

Mineral Saturation States and Aquifer Mineralogy

The producing aquifers for Ue29a#2 and the J-wells are volcanic rock. The depth and lithology of the Lathrop Wells well are unknown, but the well is reasonably expected to be in alluvium. The alluvium should be comprised predominantly of volcanic rock fragments, but there is a possibility for some carbonate clasts. Groundwater chemistry in volcanic rock is controlled by glass and diagenetic minerals (smectites, zeolites, feldspars) (Claassen and White, 1979). It is also reasonable

that there is some interaction with carbonates in the soil zone and fracture linings, with the original carbonate source presumably being aeolian dust. Chemical speciation and saturation with respect to minerals were determined using WATEQ4F (Ball and Nordstrom, 1991). The chemical formula for selected minerals and the saturation state of the groundwaters with respect to those minerals are given in the Appendix. All of the samples were at or above saturation with respect to chalcedony and quartz. Trace metal analyses were only available for J-12 and J-13 (Table 2). These waters were at saturation, and in some cases quite supersaturated, with respect to gibbsite, kaolinite, albite, potassium mica, calcium montmorillonite, and illite. All of the waters remain undersaturated with respect to dolomite, but calcite saturation is approached along the flowpath. The saturation index, or log of the ratio of the ion activity product and the solubility product ($\log \text{IAP/KT}$), for calcite is -1.39 at Ue29a#2, -0.95 at J-13, -0.67 at J-12, and -0.2 at Lathrop Wells. All of the water samples have calculated P_{CO_2} values in excess of atmospheric, with the source presumably being plant root respiration introduced to the water during infiltration through the unsaturated zone. The calculated values of P_{CO_2} change downgradient as follows ($\log P_{\text{CO}_2}$): -2.0 at Ue29a#2, -1.9 at J-13, -2.3 at J-12, and -2.8 at Lathrop Wells. With the depths of the producing intervals varying between 87 m at Ue29a#2 and 307 m at J-13 (the Lathrop Wells well producing interval is unknown), the system is assumed closed with respect to atmospheric carbon.

TABLE 2. CHEMICAL AND ISOTOPIC DATA USED FOR THE FORTYMILE WASH CALCULATIONS. All units are mg/l, unless otherwise noted.

WELL	Ue29a#2	J-13	J-12	Lathrop Wells
Source	Benson <i>et al.</i>	Robinson & Beetem	Robinson & Beetem	Claassen
Temp, °C	22.7	31.1	26.7	25.1
pH	7.0	7.1	7.4	8.0
SiO ₂	44	58	56	38
Ca	10	14	17	12
Mg	0.3	1.8	0.9	0.7
Na	44	48	40	93
K	1.3	5.0	4.6	3.9
Cl	8.8	7.4	7.4	17
HCO ₃	107	136	120	153
SO ₄	21	23	21	78
F	0.9	2.4	2.2	
NO ₃		4.5	5.3	
Al		0.062	0.035	
Ba		0.02	0.007	
B		0.14	0.11	
Cu		0.005	0.001	
Fe		0.022	0.016	
Pb		0.003	0.004	
Li	0.11	0.046	0.022	
Mn		0.028		
Mo		0.015	0.004	
Sr	0.033	0.06	0.032	
V		0.009	0.004	
$\delta^{13}\text{C}$, ‰	-13.1	-7.8	-7.9	-7.4
^{14}C , PMC	60	29.3	32.2	11.7

Sources: Benson *et al.* (1983), Robinson and Beetem (1965), and Claassen (1985).

Chemical and Isotopic Reaction Models

The final NETPATH model presented here uses carbon, calcium, magnesium, and sulfur concentrations at the upgradient and downgradient wells as constraints, and considers calcite, dolomite, gypsum, calcium montmorillonite, and magnesium montmorillonite as reacting phases. The model was calibrated against $\delta^{13}\text{C}$ values. Other elements and phases were not included because they do not impact the carbon chemistry.

The only mineral phases that have a direct bearing on the carbon isotope evolution are those that contain carbon. In the NETPATH database, the available carbonate minerals are aragonite, calcite, dolomite, magnesite, magnesium-bearing calcite, rhodochrosite, siderite, strontianite, witherite, a sodium carbonate, lignite, and gas phases for CO_2 and CH_4 . Because trace element data were limited to only the J-wells, and because the trace carbonates are not expected to play an important role in the carbonate evolution, only calcite and dolomite were used in the final simulations. The choice of the calcium carbonate phase, whether aragonite or calcite, is of no importance in the NETPATH calculations.

To constrain the carbon transfer, the evolution of the accompanying cations, calcium and magnesium, was also modeled. Calcium and magnesium concentrations essentially do not increase along the flowpath, though there is an increase in carbon reflected by the increasing bicarbonate concentrations. To increase the carbon through carbonate dissolution (consistent with the undersaturated state of the upgradient wells with respect to calcite), sinks for the cations were necessary. In the final model, the cation sinks are calcium and magnesium montmorillonite. Saturation with regard to magnesium montmorillonite is unknown, but as stated in the previous section, the J-wells (which have aluminum analyses) are supersaturated with respect to calcium montmorillonite, which is an expected diagenetic phase (Matuska, 1989) and silica concentrations decrease from the J-wells to Lathrop Wells. Another reasonable scenario, especially in light of the increase in sodium seen between the J-wells and Lathrop Wells without a concomitant increase in chloride, is that calcium produced by carbonate dissolution is exchanged onto clays, releasing sodium into solution. Simulations using the exchange scenario were also run and successfully modeled, but these simulations did not result in any significant variation of carbon isotope evolution, so are not presented for the sake of simplicity. With calcium chosen as a limiting constraint, gypsum also needed to be included because a large increase in the sulfate concentration occurs from the J-wells to Lathrop Wells.

Modeling Results

Using the constraints and phases outlined above, NETPATH found four mass-balance scenarios for each of the well pairs considered. This means that four combinations of dissolution/precipitation reactions for the given phases could produce the observed change in the four constraining ion concentrations between the initial and final wells. The four NETPATH models were then evaluated and unreasonable models were eliminated on the basis of chemical and isotopic considerations. A scenario was considered unrealistic if it used precipitation of either dolomite or gypsum, or dissolution of montmorillonite, because these reactions are inconsistent with the solubility calculations. With calcite approaching saturation along the flowpath and given uncertainty in *in situ* pH (which could alter the calculated calcite saturation state), calcite was allowed to either dissolve or precipitate. Though pH is important for determining the saturation state

of the water with respect to carbon-bearing minerals, NETPATH does not maintain a hydrogen ion balance during the calculations and thus cannot be constrained by pH. A model was also considered unrealistic if the computed $\delta^{13}\text{C}$ value differed substantially from the value measured at the final well. The standard laboratory error for a $\delta^{13}\text{C}$ measurement is $\pm 0.2\text{‰}$, so the ideal calibration would be to within 0.4‰ . Realistically, larger differences are possible because of the large variation of values reported by different laboratories for samples from the same wells.

Only one model, between one well pair, met the criteria described above. This model simulated the chemical and isotopic evolution from well J-13 to Lathrop Wells. In this model, 0.03 mmol/L of calcite are precipitated, 0.57 mmol/L of gypsum are dissolved, 3.54 mmol/L of calcium montmorillonite are precipitated, and 0.26 mmol/L of magnesium montmorillonite are precipitated. The $\delta^{13}\text{C}$ calculated by this model for Lathrop Wells is -7.8‰ , compared to the observed value of -7.4‰ . Two other models of this flowpath result in calculated $\delta^{13}\text{C}$ values of -7.7 and -7.8‰ , but both allow a very small amount of dolomite precipitation (0.04 and 0.01 mmol/L , respectively). Given the uncertainty in reaction kinetics, these models are also considered viable. The first of these allows 0.06 mmol/L of calcite dissolution, and the second does not include the calcite phase. Models of the similar flowpath from J-12 to Lathrop Wells do not agree as well because there is a larger increase in carbon, as compared to J-13 to Lathrop Wells, and this results in more carbonate dissolution and a computed $\delta^{13}\text{C}$ value heavier than measured. For example, a model with 0.42 mmol/L calcite dissolution, 0.59 mmol/L gypsum dissolution, 6.8 mmol/L calcium montmorillonite precipitation, and 0.04 mmol/L magnesium montmorillonite precipitation gives a computed $\delta^{13}\text{C}$ value of -6.6‰ , as compared to the measured -7.4‰ . This is beyond expected laboratory error, but given the possibility of larger measurement uncertainty between different laboratories, and the sensitivity of the model to assigned $\delta^{13}\text{C}$ values, the models of isotopic evolution from both J-12 and J-13 to Lathrop Wells are considered viable.

Some model uncertainty is introduced by the NETPATH treatment of bicarbonate. Similar to WATEQ4F, a number of options are available to the user for specifying carbon input. The results given above used the option of specifying field alkalinity as bicarbonate (HCO_3^-), as the data were obtained from field alkalinity titrations. NETPATH apparently adjusts the number of moles of carbon from this value, but the procedure followed is not explained in the documentation. This adjustment results in the total carbon concentration calculated by NETPATH at J-13 (2.56 mmol/L) being slightly higher than that at Lathrop Wells (2.53 mmol/L), despite J-13 having a lower field alkalinity measurement (HCO_3^- of 136 mg/L at J-13 versus 153 mg/L at Lathrop Wells). Pending an inquiry to the NETPATH authors regarding the carbon calculation, the sensitivity of the results was evaluated by using the option of specifying the number of moles of carbon, which NETPATH does not alter. This results in the dissolution of 0.3 mmol/L calcite rather than precipitation of 0.03 mmol/L as described above, but the isotopic results are not drastically altered: the modeled $\delta^{13}\text{C}$ under the dissolution scenario is -7.9‰ .

No acceptable models were found to simulate the isotopic evolution from Ue29a#2 to either J-13 or J-12. The difference in $\delta^{13}\text{C}$ between the models and the observed value at J-13 varies from 2.2 to 3.1‰ . The computed values are consistently too light in $\delta^{13}\text{C}$, predicting values of -10 to -11‰ . The disagreement is greater between Ue29a#2 and J-12, with $\delta^{13}\text{C}$ differences of 3.4 to 5.2‰ , again with the model predicting a depletion in $\delta^{13}\text{C}$. The inconsistency between the models and measured data indicate that all carbon sources and sinks have not been modeled correctly;

specifically, there is an excess of heavy carbon that cannot be attributed to carbonate rock dissolution because of the ion-concentration constraints. The solid phase $\delta^{13}\text{C}$ was assumed to be 0‰ in all scenarios, but given limited available NTS data (Boughton, 1986), it is more likely to be lighter rather than heavier in the Fortymile Wash area (caliche values ranged from -7.9 to -4.6‰). Isotopic exchange is one method of increasing the $\delta^{13}\text{C}$ without affecting the amount of ions in solution, but is difficult to justify in a non-carbonate aquifer with relatively short residence times and low temperatures. Rather, the hydrogeologic environment suggests that there is additional input of water to the system between Ue29a#2 and the J wells, possibly as infiltration of runoff in Fortymile Wash. This recharge occurs at lower elevations than recharge to Ue29a#2 and upgradient areas, and may account for the difference in $\delta^{13}\text{C}$ through a variation in the balance of C-3 and C-4 plants. The successful modeling of groundwater flow from the J wells to Lathrop Wells, without considering mixing with recharge, suggests that infiltration of runoff to the water table is less important in the lower reaches of Fortymile Wash. This is also consistent with the hydrogeologic setting as flow is reasonably expected to be less frequent in the lower reaches and subject to greater spreading with subsequent losses by evapotranspiration.

¹⁴Carbon-Based Groundwater Velocity in Lower Fortymile Wash

Without a successful model of the geochemical and stable carbon isotope evolution, the change in ^{14}C concentration attributed to radioactive decay cannot be identified. Thus, it is not prudent to calculate groundwater velocities using ^{14}C between Ue29a#2 and the downgradient wells. Velocities can be calculated from the successful models of evolution from the J wells to Lathrop Wells. NETPATH computes the difference between the amount of ^{14}C measured at the downgradient well and the amount expected to be there based on the chemical reactions invoked in the model. The difference is attributed to radioactive decay and, with the construction of these models, represents the transit time from the initial to final well. Because the carbon isotope evolution is essentially the result of dilution by rock with no modern carbon along the flowpath, the NETPATH ages can be checked against calculations using a simple $\delta^{13}\text{C}$ correction (see Table 1). The travel time from J-13 to Lathrop Wells calculated by NETPATH is 7589 years, while 7154 years result from the $\delta^{13}\text{C}$ corrected ages. The travel time calculated by NETPATH for flow from J-12 to Lathrop Wells is 6875 years, with a $\delta^{13}\text{C}$ corrected time of 7829 years. Ignoring any possible vertical components of flow, the path length from J-13 to Lathrop Wells is 17,965 m, and the length from J-12 to Lathrop Wells is 13,177 m. These parameters lead to estimates of groundwater velocity of 2.4 m/yr (NETPATH) to 2.5 m/yr ($\delta^{13}\text{C}$ corrected) from J-13 to Lathrop Wells and 1.9 m/yr (NETPATH) to 1.7 m/yr ($\delta^{13}\text{C}$ corrected) for J-12 to Lathrop Wells.

The transit time results are relatively insensitive to slight variations between the models considered. For instance, if a model allowing trace dolomite precipitation (0.007 mmol/L) is considered for J-12 to Lathrop Wells, the transit time becomes 6825 years, resulting in the same velocity of 1.9 m/yr within a reasonable rounding error. The calculations using dissolved carbon as the input, with resultant calcite dissolution rather than precipitation, gave a travel time of 7684 years and velocity of 2.3 m/yr. It is important to remember that the determination of a consistent model using reasonable phases is not proof of occurrence. In this regard, it is useful to consider groundwater evolution from J-13 to well 16s/49E-5acc (Claassen's (1985) well no. 4). This well is located directly in the Fortymile Wash drainage, west and south of Lathrop Wells. Using the same phases and constraints as the other models, the cations can only be modeled with dissolution of a

calcium-bearing phase (such as calcium montmorillonite). If the calcium-producing reaction does not affect the carbon system, the velocities calculated by NETPATH will still be valid, even though the chemical model may be incomplete. If this is true for well 16s/49E-5acc, the transit time calculated by NETPATH is 3337 years, resulting in a velocity of 7.5 m/yr. This faster velocity is a result of the higher PMC at well 16s/49E-5acc as compared to Lathrop Wells (19.3 vs. 11.7). It is impossible to determine if these differences result from inadequacies in the model or natural spatial variability, especially related to proximity to the channel. Given the consistency of the Lathrop Wells models with the equilibrium thermodynamics of the J wells (the only wells for which the state of the clay minerals could be determined), and the closeness of the stable carbon isotope fit, more confidence is suggested in those velocities, but other Amargosa Valley data should be evaluated. For comparison, White and Chuma (1987) used EQ3NR/EQ6 to model chemical reactions for a large number of Amargosa wells and calculated a range of flow rates between 3 and 30 m/yr between J-13 and the north central Amargosa Desert.

FLOWPATH IN THE CARBONATE AQUIFER OF NORTHERN YUCCA FLAT

Mineral Saturation States and Aquifer Mineralogy

Both TW2 and Ue1q are completed in the Paleozoic Lower Carbonate Aquifer, primarily composed of limestone and dolomite. The TW2 producing unit is generally described as a silty and brecciated dolomite. The abridged log lists argillite interbedded with primarily dolomite and minor limestone, and there are notes of a few marble zones (Moore *et al.*, 1963). The limestone and dolomite at Ue1q is also noted as being argillaceous, with iron staining, and x-ray diffraction of drill cutting samples collected during recompletion revealed the fines to consist in large part of clinoptilolite and alkali feldspar, with minor montmorillonite and illite muscovite (Drellack, 1992). WATEQ4F (Ball and Nordstrom, 1991) was used to compute the mineral saturation indices (Appendix). Both wells were found to be at saturation with respect to calcite, dolomite, quartz, and chalcedony and undersaturated with respect to gypsum, while TW2 was also supersaturated with respect to adularia, gibbsite, kaolinite, and laumontite (aluminum data were not available for Ue1q). The dissolved CO₂ was well above atmospheric, with a log P_{CO2} = -2.28 at TW2 and a log P_{CO2} = -2.57 at well Ue1q.

Chemical and Isotopic Reaction Models

A water chemistry analysis from Robinson and Beetem (1965) was used for TW2, since this was the only source with field measurements of pH and alkalinity. Laboratory measurements of pH and bicarbonate ion from this well were about 10% higher than those measured in the field. Two carbon isotope analyses were available for TW2. The sample chosen for this simulation was collected by Boughton (1986); this sample was about 4‰ heavier in δ¹³C and 5 PMC less than the sample reported by Grove *et al.* (1969).

Field parameters were measured by DRI at well Ue1q after the well was recompleted in 1992; the chemical analysis and field measurements from 1992 were used for this well. Two carbon isotope analyses were also available from the post-recompletion well Ue1q. The first sample was collected by DRI using a pump during aquifer testing, the second sample was collected by LLNL via a discrete sampler. The DRI data are used in the calculations.

Mass transfer was calculated with NETPATH using TW2 as the initial well and well Ue1q as the final well. The major carbonate mineral phases were assumed to be calcite and dolomite, and gypsum and calcium montmorillonite were included as sources and sinks of available calcium ion. Carbon, calcium, magnesium, and sulfate were used as the constraining ions. Using this combination of constraints and phases resulted in one model solution. The initial model suggested that some dead-rock carbon exchange was occurring along the flowpath, because the observed $\delta^{13}\text{C}$ values were heavier than the computed values (computed $\delta^{13}\text{C}$ of -10.8‰ versus -8.5‰ observed). Isotope exchange is considered reasonable on this flowpath given the carbonate nature of the aquifer, slightly elevated temperatures (34°C), and relatively long residence time for groundwater. The final model invokes 0.22 mmol/L dissolution of dolomite, 0.48 mmol/L dissolution of gypsum, 0.18 mmol/L precipitation of calcite, and 3.17 mmol/L precipitation of calcium montmorillonite, with 0.6 mmol/L carbon isotopic exchange between calcite and groundwater. The $\delta^{13}\text{C}$ of the calcite carbon was assumed to be 0‰ . Both the computed and observed $\delta^{13}\text{C}$ values were -8.5‰ . This match indicates that most sources and sinks of carbon along the flowpath were included in the model. However, there is uncertainty as to the sink for calcium ions. Given the argillaceous nature of the aquifers noted at each well, there are any number of clay minerals and zeolites that might precipitate or exchange calcium. If a calcium sink other than calcium montmorillonite is used, the carbon isotope calculations will not necessarily be affected, but the balance of dolomite, calcite and exchange reactions could shift slightly to accommodate the constraints placed by the magnesium concentrations. The net effect is considered small given other sources of uncertainty (discussed below).

^{14}C -Based Groundwater Velocity in Northern Yucca Flat

The NETPATH modeling results suggest that ^{14}C can be used to calculate flow velocities between TW2 and Ue1q. NETPATH computes the difference between the amount of ^{14}C measured at the downgradient well and the amount expected to be there based on the chemical reactions invoked in the model. The difference is attributed to radioactive decay, and with the construction of this model, represents the transit time from the initial to final well. Because the carbon isotope evolution in this model is essentially the result of dilution by dead rock carbon along the flowpath (as in the Fortymile Wash cases), the NETPATH age can be checked against calculations using a simple $\delta^{13}\text{C}$ correction. The travel time from TW2 to Ue1q calculated by NETPATH is 5392 years, while 5719 years result from the $\delta^{13}\text{C}$ -based correction. Ignoring any possible vertical components of flow, the path length from TW2 to Ue1q is 9841 m, leading to a NETPATH-based estimate of groundwater velocity of 1.8 m/yr.

The single largest uncertainty in the velocity estimate is the issue of whether TW2 and Ue1q are along a carbonate flowpath, or are instead on parallel flowlines toward the axis of Yucca Flat. Confirmation of the hydraulic head relationship and examination of the flow directions predicted by the numerical model should resolve this concern. Additional uncertainty stems from uncertainty in the carbon isotope data, rather than in the chemical model. The difference in $\delta^{13}\text{C}$ between DRI and LLNL is large and needs to be resolved. The models presented would have to be substantially altered if calibrated to the LLNL data. In addition, rock carbon isotopic data are unavailable for TW2 or Ue1q, and those values are important in the calculations of isotopic exchange. $\delta^{13}\text{C}$ analysis of limestone and dolomite rock cores from the NTS yielded a range from $+4.1$ to -1.9 (Boughton, 1986). For instance, increasing the rock $\delta^{13}\text{C}$ from 0 to $+2\text{‰}$, results in less exchange, less dilution

by dead ^{14}C , and travel times higher by several hundred years. For comparison, Winograd and Thordarson (1975) used hydraulic parameters to estimate a range of flow velocities in the carbonate aquifer of central Yucca Flat of 1.8 to 183 m/yr, depending on the effective fracture porosity assumed. Grove *et al.* (1969) calculated a velocity of 2.1 m/yr from TW2 to Well C-1 based on ^{14}C , though as discussed previously, the Yucca Fault hydraulic barrier crosses that hypothetical flowpath.

FLOWPATH IN THE CARBONATE AQUIFER OF SOUTHERN YUCCA FLAT

Winograd and Thordarson (1975) evaluated the hydrology of Yucca Flat. Their conceptual model described a relatively flat water table in the Lower Carbonate Aquifer underlying the volcanic testing zone and delineated groundwater moving southward from central Yucca Flat toward Yucca Lake. They also suggested that some downward movement of groundwater from the volcanic rocks to the Lower Carbonate Aquifer may be occurring. A general southward flowpath in the Lower Carbonate Aquifer from ER-6-1 to Well C-1 was selected for further geochemical modeling. In general, major ion concentrations increase from ER-6-1 to Well C-1 and carbon isotopes become progressively heavier along the proposed flowpath (Table 3). Trace-metal analyses were available for both wells and were used to evaluate chemical speciation. The proposed flowpath is assumed to be in a closed system relative to the unsaturated-zone carbon dioxide reservoir.

TABLE 3. CHEMICAL AND ISOTOPIC DATA USED FOR THE YUCCA FLAT FLOWPATHS. All units are mg/l, unless otherwise noted.

WELL	Test Well 2	Ue1q	ER-6-1	Well C-1
Source	Robinson & Beetem	DRI	DRI	Robinson & Beetem
Temp, °C	33.9	31.5	43.7	36.1
pH	7.6	7.73	7.31	6.9
SiO ₂	46	50.5	34.3	29
Ca	30	28.8	33.4	74
Mg	10	14.8	12.2	29
Na	28	29.5	47.0	123
K	6.6	5.0	8.32	14
Cl	5.6	6.3	11.8	34
HCO ₃	186	205	228	574
SO ₄	21	22.1	32.1	66
F	0.5	0.47	0.68	1.2
NO ₃	4.1	4.21	0.58	0.0
Al	0.036	<0.02	<0.02	0.064
Ba	0.01	0.058	0.134	0.12
B	0.19			0.53
Cu	0.007	0.022	0.034	0.098
Fe	0.008	0.01	0.06	0.02
Pb	0.005	<0.001	<0.001	0.028
Li	0.01	0.021	0.035	0.24
Mn		<0.005	0.009	
Mo	0.0008			
Sr	0.036			0.6
V	0.002	<0.02	<0.02	
$\delta^{13}\text{C}$, ‰	-11.2	-8.5	-6.3	-3.8
^{14}C , PMC	10	3.8	2.4	0.8

Chemistry Sources: Robinson and Beetem (1965), and Desert Research Institute files. Isotope data sources are given in Table 1.

Mineral Saturation States and Aquifer Mineralogy

Both ER-6-1 and Well C-1 are completed in the Paleozoic Lower Carbonate Aquifer, which is composed primarily of limestone and dolomite (Winograd and Thordarson, 1975; Gillespie 1993). Because of high-angle, normal, extensional faulting, flow in Yucca Flat may be highly compartmentalized. This type of geologic terrain may result in groundwater in the carbonate aquifer encountering zones of volcanic rocks which may change the chemistry and possible mineral phases; although no readily apparent structure is known to be present along the proposed flowpath. The equilibrium thermodynamic speciation code WATEQ4F (Ball and Nordstrom, 1991) calculated the groundwater at ER-6-1 to be saturated with respect to calcite, dolomite, barite, chalcedony, and quartz; undersaturated with respect to gypsum, rhodochrosite, and witherite; and to have a calculated $\log P_{CO_2}$ of -1.82. Equilibrium calculations for groundwater from Well C-1 suggest saturation with respect to calcite, dolomite, barite, and chalcedony; undersaturation with respect to gypsum, albite, analcime, anorthite, strontianite, and witherite; and supersaturation with respect to adularia, beidellite, gibbsite, illite, kaolinite, laumontite, leonhardite, calcium montmorillonite, phillipsite, and pyrophyllite; and a calculated $\log P_{CO_2}$ of -1.099. The saturation indices are summarized in the Appendix, along with the mineral formulas.

Chemical and Isotopic Reaction Models

The NETPATH model presented uses calcium, magnesium, sodium, carbon, and sulfur as constraints and considers calcite, dolomite, gypsum, calcium montmorillonite, sodium montmorillonite, analcime, and cation exchange as reacting phases. The model was calibrated against $\delta^{13}C$. Other elements and phases were not included in the final model because they do not affect the carbonate chemistry and therefore the $\delta^{13}C$ calibration. It was assumed that all dissolving carbonate mineral phases had a $\delta^{13}C$ signature of 0.0‰. Because of the great depth to water and the large thickness of overlying saturated volcanic rock, the system was considered to be closed to any carbon dioxide reservoir.

As mentioned above, all major ion concentrations increase from ER-6-1 to Well C-1 with carbon increasing dramatically from 4.0 to 11.4 mmol/L. Because of this large increase in carbon, both calcite and dolomite are modeled to dissolve along the flowpath to produce the necessary final carbon concentration. The amounts of calcite and dolomite dissolved are constrained by the change in calcium and magnesium. The increase in sulfur requires the dissolution of gypsum which produces more calcium. Finally, sodium was used to constrain both the dissolution of sodium-bearing phases sodium montmorillonite and analcime and to consider calcium-sodium cation exchange.

Modeling Results

Using the constraints and phases listed above, NETPATH produced four models for the proposed flowpath. All four models include dissolution of 6.0 mmol/L of calcite, 0.7 mmol/L of dolomite, and 0.35 mmol/L of gypsum. All four models have the same final computed $\delta^{13}C$ value for Well C-1 (-2.22‰) and the same flowpath age of 470 years (0.85 PMC calculated vs. 0.8 PMC measured). Variations between the four models result from different ways of handling the final sodium and calcium concentrations and do not affect the carbon isotope calculations. Model variations include the dissolution of analcime or sodium montmorillonite, the precipitation of calcium montmorillonite, and cation exchange.

The modeled variations in the chemical reactions principally revolve around accounting for the final calcium concentration. Because of the carbon and sulfur constraints, much more calcium is produced than indicated by the water chemistry data; therefore, calcium must be removed by some mechanism. In model one, 36 mmol/L of calcium montmorillonite are precipitated and 10 mmol/L of sodium montmorillonite are dissolved. In model two, 36 mmol/L of calcium montmorillonite are precipitated and 3.3 mmol/L of analcime are dissolved. In model three, 26 mmol/L of calcium montmorillonite are precipitated and 1.65 mmol/L of calcium-sodium exchange occur. In model four, 26.7 mmol/L of sodium montmorillonite are dissolved and 1.65 mmol/L of calcium-sodium exchange occur.

WATEQ4F calculations suggest that precipitation of calcium-bearing secondary minerals may be occurring. Laumontite ($\log \text{IAP/KT} = 2.229$), leonhardite (12.009), and calcium montmorillonite (3.705) are all supersaturated at Well C-1 and can be considered possible precipitating calcium phases. Both laumontite and leonhardite are calcium zeolites and are possibly prevalent in the overlying volcanic rocks. Because NETPATH does not include these minerals in its database, most of the excess calcium was attributed to precipitation of calcium montmorillonite in the model. WATEQ4F is less definitive with secondary sodium phases. The thermodynamic calculations show supersaturation with respect to the sodium montmorillonite beidellite ($\log \text{IAP/KT} = 3.898$) and the sodium-bearing zeolite, phillipsite (1.298). Again, these phases are not in the NETPATH database, so sodium montmorillonite was used to account for any sodium phases. Because equilibrium thermodynamic calculations suggest the presence of zeolitic minerals, some calcium removal may also reasonably be attributed to calcium-sodium cation exchange.

Well C-1 has a measured $\delta^{13}\text{C}$ value of -3.8‰ and 0.8 PMC. Modeling, using an initial measured $\delta^{13}\text{C}$ value of -6.3‰ and 2.4 PMC at ER-6-1, produced a calculated $\delta^{13}\text{C}$ value of -2.22‰ and 0.85 PMC for Well C-1. This 1.6‰ difference in $\delta^{13}\text{C}$ results from addition of 0.0‰ carbonate from calcite and dolomite dissolution by Rayleigh fractionation. Assuming that analytical results are correct, this difference may be the result of less dissolution of carbonate minerals than the model suggests, or mixing of another groundwater with lighter $\delta^{13}\text{C}$ and more modern carbon, or a $\delta^{13}\text{C}$ for carbonate minerals other than 0.0‰ . Assuming that the chemistry balance is correct and the carbon cannot be coming from another mineral source, then either mixing or a carbonate mineral $\delta^{13}\text{C}$ other than 0.0‰ is likely. A preliminary attempt to match the calculated and measured $\delta^{13}\text{C}$ at Well C-1 by mixing with a volcanic aquifer water (from Test Well B) did not produce any reasonable results. This does not rule out the possibility, but because of time constraints, a detailed examination of this scenario is beyond the scope of this project. Measured $\delta^{13}\text{C}$ signatures of carbonate rocks in the NTS area are very limited. Boughton (1986) conducted eight analyses of NTS carbonate rock core. Her $\delta^{13}\text{C}$ values ranged between -1.9 and $+4.1\text{‰}$ and were from UE-16d, Army Well 1, Army Well 2, U-3cn#5, and Well C. The Well C value was -1.9‰ . This value was input into the model instead of 0.0‰ and resulted in a calculated $\delta^{13}\text{C}$ of -3.2‰ for Well C-1 which is much closer to the measured value; however, this exercise did not affect the final calculated PMC for Well C-1 because the number of moles of carbonate minerals which dissolve do not change and these minerals still have 0 PMC.

¹⁴C-Based Groundwater Velocity in Southern Yucca Flat

Using the NETPATH calculated flowpath age of 470 years and a distance of 7449 m straight-line flowpath (based on Nevada coordinates), a groundwater velocity of 15.9 m/year results. This velocity is somewhat larger than other velocities calculated in this report but is within the range of Yucca Flat velocities reported by Winograd and Thordarson (1975) for the Lower Carbonate Aquifer (1.8 to 183 m/yr). Uncertainty in the NETPATH-based velocity is related primarily to uncertainty in whether the model accounts for all carbon sources and sinks properly, because of the discrepancies in modeled vs. measured $\delta^{13}\text{C}$ values. It is likely that the groundwater at Well C-1 is receiving a small component of downward moving, slightly younger volcanic aquifer water (for example Test Well B, $\delta^{13}\text{C} = -10.7$, PMC = 20.3) which may be slightly increasing the measured PMC.

CONCLUSIONS

Chemical and isotopic data from wells at the NTS provide a powerful tool for verifying the predictions of complex numerical flow models. Three well transects along possible flowpaths were identified for quantitative analysis in this study: Ue29a#2 to J-13 to J-12 to Lathrop Wells along Fortymile Wash, Test Well 2 to Ue1q in northern Yucca Flat, and ER-6-1 to Well C in southern Yucca Flat. The chemical evolution from Ue29a#2 to downgradient wells could not be successfully modeled as a closed system. The chemical and isotopic evolution from upgradient to downgradient wells was modeled for the other well pairs using carbon, sulfate, calcium, and magnesium as constraints and calibrating to measured $\delta^{13}\text{C}$ values. The resultant models (Table 4) are consistent with equilibrium thermodynamic reactions for the phases considered. Assuming that all important constraints and phases were included in the models, groundwater flow velocities can be calculated from the ¹⁴C-based transit times between the initial and final wells. Groundwater flow velocities in the fractured volcanic aquifer beneath lower Fortymile Wash are estimated to range from 1.9 to 2.4 m/yr. Velocities in the carbonate aquifer of northern Yucca Flat are estimated to be 1.8 m/y, while the estimated velocity in the carbonate aquifer of southern Yucca Flat is higher at 15.9 m/yr.

Though the chemical models derived here are consistent with available data and thermodynamics, that is no guarantee that they accurately model the system. Confirmation of the flowpaths selected is needed from the hydraulically-based groundwater flow model, particularly for the path from TW2 to Ue1q. Uncertainty in the chemical model would be greatly reduced with mineralogic information from the aquifers and trace metal analyses of groundwater from the wells. Stable carbon isotope data are also needed for the reacting solid phase in the aquifer, and could be provided by analysis of cores. In some cases, additional carbon isotopic data from groundwater samples are also needed to reduce data discrepancies. Chemical and isotopic data are available for many more flowpaths at the NTS, but they are not considered to be closed systems amenable to the simple mass-balance models presented here. Analysis of these more complex evolutionary paths could provide important velocity estimates, as well as information regarding mixing between aquifers.

TABLE 4. SUMMARY OF NETPATH MODELING RESULTS PRESENTED IN THE TEXT. The phases listed were dissolved or precipitated as indicated to achieve the ion mass-balance constraints. The initial and final constraint concentrations are those measured in water from the beginning and ending wells. The difference between the observed and computed ^{14}C concentrations yields the travel time.

	J-13 to Lathrop Wells concentration, mmoles/L		J-12 to Lathrop Wells concentration, mmoles/L		TW2 to Ue1q concentration, mmoles/L		ER-6-1 to Well C-1 concentration, mmoles/L	
	initial	final	initial	final	initial	final	initial	final
constraints								
calcium	0.35	0.30	0.42	0.30	0.75	0.73	0.83	1.85
magnesium	0.07	0.03	0.04	0.03	0.41	0.63	0.50	1.19
carbon	2.56	2.53	2.11	2.53	3.17	3.42	4.04	11.5
sulfur	0.24	0.81	0.22	0.81	0.22	0.70	0.33	0.69
phases	mmoles/L		mmoles/L		mmoles/L		mmoles/L	
	calcite	0.03 ppt.	0.42 diss.		0.18 ppt.		6.0 diss.	
	dolomite	—	—		0.22 diss.		0.7 diss.	
	gypsum	0.57 diss.	0.59 diss.		0.48 diss.		0.35 diss.	
	Ca-mont	3.54 ppt.	6.79 ppt.		3.17 ppt.		36 ppt.	
	Mg-mont	0.26 ppt.	0.04 ppt.		—		—	
	Na-mont	—	—		—		10 diss.	
	isotope exchange	—	—		0.6		—	
	observed	computed	observed	computed	observed	computed	observed	computed
^{13}C , ‰	-7.4	-7.8	-7.4	-6.6	-8.5	-8.5	-3.8	2.2*
^{14}C , PMC	11.7	29.3	11.7	26.9	3.8	7.3	0.8	0.85
travel time, years	7589		6875		5392		470	
path length, m	17,965		13,177		9841		7449	
velocity, m/yr	2.4		1.9		1.8		15.6	

*The computed $\delta^{13}\text{C}$ value is -3.2‰ if the rock $\delta^{13}\text{C}$ is -1.9‰ (as measured in Well C core; Boughton, 1986), instead of 0‰ .

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