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**COMMENTARY BY JERRY S. SZYMANSKI AND C. B.
ARCHAMBEAU REGARDING "SPRING DEPOSITS AND LATE
PLEISTOCENE GROUND-WATER LEVELS IN SOUTHERN
NEVADA", BY J. QUADE.**

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**DIALOGS BY JERRY S. SZYMANSKI AND C. B. ARCHAMBEAU REGARDING
"SPRING DEPOSITS AND LATE PLEISTOCENE GROUND-WATER LEVELS IN
SOUTHERN NEVADA"**

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Review of the paper:

**Spring Deposits and Late Pleistocene Ground-Water Levels in Southern
Nevada**

by Dr. Jay Quade of the University of Arizona

Introduction

Proceedings of the Fifth Annual International Conference on High Level Radioactive Waste Management include a paper written by Dr. J. Quade and entitled *Spring Deposits and Late Pleistocene Ground-Water Levels in Southern Nevada*. A copy of this paper is provided in the attached Appendix A. The paper is concerned with "...the maximum elevation attained by water tables in the region during the Quaternary. This information is crucial to establishing the feasibility of a repository for high-level nuclear waste at Yucca Mountain." Specifically, this information may be used to evaluate magnitudes of the anticipated changes of the local water table and, based upon it, overall impact of these changes on long-term performance of the proposed high-level nuclear waste repository.

The work "was funded mainly by the Nevada Nuclear Waste Project Office." As such it may be used, by the DOE scientific establishment, to further the ongoing misrepresentation of hydro-tectonic setting of Yucca Mountain. With this possibility in mind, TRAC-NA performed an in-depth review of Quade's paper. The results of

this review are presented below and are concerned with the following three subjects:

- a) misleading information and erroneous interpretations; b) omission of relevant and available data and observations; and c) reasonable estimates of the anticipated rises of the water table at Yucca Mountain.

Misleading Information and Erroneous Interpretations

The paper by Quade (1994) presents some misleading information and some erroneous interpretations. In that regard, the following three remarks seem appropriate.

First, by stating on p. 2533 "...our comparison reveals that ground water levels have witnessed only modest changes (Δ WT) over the past ~17,000 years, from 5 to at most 115 m (Table 1), far short of the > 500 m estimates of others" Quade misleads a reader with regard to the opinions expressed by Szymanski (1989). These opinions are summarized in the attached Appendix B. Rather than advocating the >500 m change in the water table, Szymanski expressed a considerable uncertainty in that regard. Specifically, in the 1989 report he stated: "The above crude estimates indicate, that at Yucca Mountain, magnitudes of the long-term displacements of the water table involve several tens and perhaps a few hundred meters. Displacements of the water table of such magnitudes are likely to profoundly impact the long-term performance of the proposed high-level nuclear waste repository. Also, in extreme cases such displacements may result in flooding of the repository and expulsion of ground water at the land surface. A critical point in

the overall conceptual considerations of the Yucca Mountain ground water system is to establish unequivocally whether or not the geologic record representing Plio-Quaternary time contains expressions of the past large-scale displacements of the water table. As discussed in Section 4.5.3, there are indications that this may be the case indeed. Thus far, however, only superficial studies of the calcite-opaline silica vein deposits were performed during the last 10 years. Consequently, the current data base does not allow for an unequivocal resolution of this critical issue. Nevertheless, the possibility that the veins constitute a record of large-scale displacements of the water table in the past is real and, by all means, not remote."

Second, by stating on p. 2534 "...the present water table is about 115 m below the highest of the Lathrop Wells deposits on southern Crater Flat, implying the full glacial (FG) rise of the water table was less than that in this area." Quade misrepresents the actual field data and observations. From the attached Fig. 1, it is evident that subject deposits occur at an altitude ranging from 800 to 850 m. As shown in Fig. 2, for the corresponding area, the contemporary water table is situated at an altitude ranging from 716 m (2350 ft) to 710 m (2330 ft). Rather than <115 m, the Crater Flat spring deposits imply a minimum of 134 m rise of the water table instead.

In addition to the actual magnitude, Quade further misleads a reader by implying that we know with certainty the reasons for the abnormally high stand of the local water table. All that is known with certainty, however, is that this high stand occurred

during the last full-glacial period when ground water recharge was relatively high. In view of that, it is permissible (but no more) to postulate that the climatic change could have been the sole factor. However, considering both the young age of local faulting and volcanism and the "...alkaline, at times slightly saline" environment of formation of the subject deposits (Hoover, et al., 1981), it is equally permissible to postulate that other factors could have also been involved (e.g. hydro-tectonic activity associated with a volcanic event and/or fault rupture). Understanding of processes which are expressed by the Crater Flat deposits is crucial to establishing a proper conceptual understanding of the Yucca Mountain hydrosphere and, based upon it, "...the feasibility of a repository for high-level nuclear waste at Yucca Mountain." By implicitly regarding plausible as certain, Quade bypasses a need for developing such understanding.

Third, by stating on p. 2534 "...the proposed repository horizon remained 85-285 m above the regional water table throughout much of the Quaternary." Quade draws an unsupported and misleading conclusion. Putting aside a number of problems [i.e.: a) the questions of true magnitude of the water table rise in the southern Crater Flat area; b) the ~20 ka age of the deposits; c) the questions regarding processes involved in producing the rise; and d) omission of relevant and available data and observations], this conclusion is based on the stated assumption "...the increased flow cross-sections, and not increases in water table gradients, would be the principal factor responding to enhanced recharge." Field observations, made down-gradient of the Crater Flat deposit outcrops, demonstrate that this assumption is not valid. In this regard, the following two observations are particularly relevant, as

illustrated in Fig. 3. On the one hand, down-gradient of the outcrops, the contemporary topographic gradient (~ 4.4 m/km, Fig. 1 and 3) is greater than the corresponding potentiometric gradient (1.35 m/km, Fig. 2 and 3). Consequently, the water table depth diminishes with the increasing down-gradient distance away from the Crater Flat deposits. On the other hand, between Highway #95 and the California-Nevada State Line, diatomaceous deposits similar to those at the southern end of Crater Flat are absent. This observation implies that during the formation of the Crater Flat deposits, the water table in the down-gradient area was situated below the topographic surface. Such a relative position of the water table may only be accounted for by assuming, at and upstream of the Crater Flat deposits, the potentiometric gradient was steeper than the topographic gradient, i.e. in excess of 4.4 m/km. In other words, the paleo-potentiometric gradient must have been 325 percent steeper than the corresponding contemporary gradient, in direct conflict with Quade's assumption. Of course, we do not know how areally extensive the increased potentiometric gradient was. Consequently, extrapolation of the observed hydrologic effects, from southern Crater Flat to Yucca Mountain, may not be made with confidence. Assumption of the climatic change as a sole factor responsible for the observed rise, however, leads to a conclusion: the corresponding water table rise at Yucca Mountain could have been as high as 222 m ($134 \text{ m} + 4.4 \text{ m/km} \times 20 \text{ km}$).

Omission of Relevant and Available Data and Observations

With reference to completeness, Quade's paper is disappointing as the surficial data

is incomplete and the subsurface data is completely ignored.

Surficial Data and Observations

With regard to the surficial data, an outstanding example of data omission is the so called Whamonie deposit. It consists of a mound, a few meters thick, composed of gypsum, calcite and sepiolite. The surficial outcrop of this mound is situated about 6 km north-west of Skull Mountain and only ~15 km north-east of Yucca Mountain. The Whamonie mound occurs in a topographic setting that precludes playa deposition and in an area underlain by a vadose zone about 500 m thick. The attached Appendix C presents some basic information regarding the Whamonie mound. Careful review of this material reveals that, in the opinion of researchers from both the United States Geological Survey and the Los Alamos National Laboratory, the Whamonie deposits represent a paleo site of hot water discharge. A USGS letter to the DOE describes the Whamonie mound as "...local surface deposits from recent warm springs" which indicate "...upward seepage of ground water, possibly from great depths." (USGS, 1980). Because both the ground water origin and the Quaternary age of the Whamonie mound are fairly certain, one must regard the statement by Quade from p. 2530 "... water levels have varied by <115 m, and in most cases much less, in the latter half of the Quaternary", with a great deal of skepticism.

Both the Whamonie mound and the southern Crater Flat spring deposits are

important for a number of reasons. Besides providing convincing arguments in support of expectations of fairly large changes in the water table, these deposits offer an opportunity for establishing reliability and scientific validity of the approach employed by Quade and Cerling (1990) in studies of the genesis of the controversial veins at Yucca Mountain. These veins are exposed in Trench 14, at an altitude some 500 m above the contemporary water table. From Fig. 4 and 5, it may be observed that the isotopic character of carbon and oxygen incorporated in ground water deposits, at the Whamonie and Crater Flat sites, are identical to those incorporated in surficial cobble encrustations. These encrustations were regarded by Quade and Cerling (1990) as deposits for which *per descensum* origin is known with certainty. Thus, they were employed as a suitable reference for judging the origin of the Trench 14 veins. The evident isotopic similarity of the deposits with supposedly differing origins, shown in Fig. 3 and 4, requires one of the following two statements to be true: either (a) *per ascensum* processes yield carbonate deposits which, in terms of stable isotope signatures, are identical to those produced via *per descensum* (pedogenic) processes, or (b) the assumption of the *per descensum* (pedogenic) origin for the reference cobble encrustations is incorrect. In either case, the reliability and validity of the methodology employed by Quade and Cerling (1990), together with their conclusions regarding the origin of the Trench 14 veins, are doubtful at best.

Another example of the data omission is a 500,000 years stable isotopic record from Devils Hole as reconstructed by Coplen et al., 1994. With reference to the contributions made by Quade (1994) and Quade and Cerling (1990), this record has

profound implications. Because the following discussions pertain to the conclusions drawn by Quade and Cerling with regard to the genesis of the controversial Trench 14 veins, copies of the papers by Coplen et al. (1994) and by Quade and Cerling (1990) are provided in the attached Appendix D.

The Devils Hole isotopic record and the essence of reasoning developed by Quade and Cerling are shown in Fig. 6. With regard to the latter, two points are worth noting. First, relative to "modern soil carbonates in the area" (or the reference cobble encrustations), the stable isotopic signatures of the controversial veins are **CONCORDANTLY** shifted toward "lighter" regions. Second, the isotopic composition of the Trench 14 veins is similar to that of the reference carbonates at higher altitudes, in the lower pinyon-juniper-sagebrush zone. This similarity was regarded by Quade and Cerling as demonstrating conclusively "...that the fracture carbonates are pedogenic in origin and that they likely formed in the presence of vegetation and rainfall typical of a glacial climate." Having thus resolved the critical safety issue, the authors concluded "...the regional water table remained below the level of Trench 14 during the time that the carbonates and silica precipitated, a period probably covering parts of at least the last 500,000 years."

This conclusion was enthusiastically welcomed by the DOE scientific establishment, which promptly heralded it to the scientific community at large and to the public. With the main obstacle out of the way, the establishment was free to proceed with the massive site characterization effort at Yucca Mountain, costing in

excess of \$1 million a day. Four years later, the establishment is prepared to continue in this investigative mode, spending several billions of dollars while exploring fine intricacies of fluid flow through a vadose zone.

But is the conclusion drawn by Quade and Cerling valid? An unequivocal answer to this question is provided in Fig. 6. With reference to southern Great Basin, this figure shows that during the last 500,000 years the carbon isotopic signature of fluids discharging from the Devils Hole cavern was **INVERSELY** related to the $\delta^{18}\text{O}$ signature. Similar anti-correlation was found in a 25 cm thick calcitic crust from Jewel Cave of South Dakota, Dublyansky (1994). Because the variations of the stable isotopic signatures in the calcitic crusts are reflecting variations in the stable isotopic compositions of soil fluids in the principal recharge areas, the Devils Hole record provides clear testimony that during periods of cool climate when atmospheric precipitation was depleted in ^{18}O the infiltrating soil fluids were enriched in ^{13}C . Although one of the stable isotopic signatures from the Trench 14 veins may be matched with the corresponding signature from "...modern soil carbonates in the area" by invoking differing climatic conditions, the remaining and climate change corrected signature is markedly different, Fig. 6a. This indicates that the differing stable isotopic signatures of the controversial and reference deposits may not possibly be attributed to precipitations under differing climatic conditions, in direct conflict with the Quade and Cerling assumption. If the reference carbonates were indeed produced via *per descensum* mechanism then it follows, the controversial Trench 14 veins must have been produced via some other mechanism. However, it is also possible that the strictly *per descensum* origin of the reference soil

carbonates was incorrectly presumed by Quade and Cerling. These carbonates may be in fact pedogenic (formed in a soil cover) derivatives of calcite-opal-sepiolite veins and breccia cements, which are ubiquitous at Yucca Mountain. One may presume that, as these derivatives became increasingly more distant (in either space or time) from their parents, they also became **CONCORDANTLY** enriched in heavy isotopes of carbon and oxygen, chiefly via evaporative enrichment mechanism. The assumption of the parent → derivative relationship is not in conflict with the Devils Hole record and satisfactorily accounts for the stable isotopic compositions of the controversial veins and the reference soil carbonates. In consideration of the above remarks, it seems prudent for Dr. Quade to revisit the question of the origin of the Trench 14 veins and reconsider accordingly the statement on p. 2534 "...the proposed repository horizon remained 85-285 m above the regional water table throughout much of the Quaternary."

Subsurface Data

With reference to the subsurface data, an outstanding example of the data omissions is a set of data concerning veins which, at Yucca Mountain, were found throughout both the vadose zone and the phreatic zone. The attached Appendix E provides a summary of this set. It contains data regarding: a) thermal environment of deposition of the veins; b) contemporary geothermal environment; c) isotopic character of carbon and oxygen incorporated in the veins; and d) U-series ages of the veins. The data is also summarized in Fig. 7 through 12.

At Yucca Mountain, the veins occur as two isotopically distinct facies. As shown in Fig. 8 through 12, one of the facies is represented by the vein set I. The corresponding veins contain isotopically "heavy" carbon and "light" oxygen. Spatially, this facies is associated with an epigenetic assemblage of alteration minerals consisting of : a) higher grade montmorillonitic clays, such as alleverdite, kalkberg, and discrete illite; b) predominantly alkalic low grade zeolites, such as clinoptilolite and mordenite; and c) higher grade minerals from the alkali series, such as analcime and albite, Bish (1989) and Szymanski (1992). The alteration minerals, from both the smectite-illite series and the zeolitic series, carry the same K/Ar ages ranging from 11.0 to about 9.5 Ma. Therefore, this alteration assemblage, and the spatially associated veins, represent hydrothermal metamorphism supported by the Timber Mountain Caldera heat source. For the purposes of this review, we do not need to be concerned with it.

The second isotopic facies consists of veins from the vein set II, Fig. 8 through 12. These veins contain isotopically "light" carbon and "heavy" oxygen. They are known to occur throughout the vadose zone as well as in the upper 700 m of the phreatic zone, for a total depth of about 1.3 km. Spatially, the second facies is associated with an epigenetic assemblage of alteration minerals consisting of: a) low grade zeolites from the alkaline earth series, such as heulandite, erionite, chabazite, stellerite; and b) higher grade alkaline earth zeolites, such as laumontite and wairakite (analcime). The corresponding alteration minerals (heulandite) carry K/Ar ages ranging from 8.5 to as little as 2.0 Ma, Wolde Gabriel (1991) and Szymanski

(1992). Importantly, for some veins representing the second isotopic facies, the mineral formation age is known independently, based on isotopic character of incorporated thorium and uranium. As shown in Appendix E (Fig. E-20 and E-21), the U/Th ages range from >400 to as little as 26 ± 2 ka. Partially, the second isotopic facies represents the Quaternary time span and as such need to be considered in reconstructing "...the maximum elevation attained by water tables in the region during the Quaternary." In view of the conclusions drawn by a number of researchers associated with the DOE establishment, one would be tempted to ignore the subject veins on the basis of their alleged *per descensum* pedogenic origin. However, there are three independent lines of evidence which are in conflict with this approach.

First, values of geothermal gradient, reconstructed based on established rate of decrease of the $\delta^{18}\text{O}$ ratio as a function of depth, range from 34 to as much as $140^\circ\text{C}/\text{km}$, (Fig. 7). For the most ^{18}O enriched veins (the isotopically "heaviest" specimens from the controversial vein set II), the inferred depositional temperatures are at least 50 percent higher than the corresponding contemporary temperatures. The significance of comparing the paleo and contemporary geothermal gradients is that any reliably established discrepancy is in conflict with the postulated *per descensum* origin of the subject veins. Reconstructed paleo-geothermal gradients from such veins ought to be equal to contemporary gradients. This is because it is totally unreasonable to associate infiltrating rainwater with a noticeable warming of rock through which this rainwater percolates. For fluids ascending from below the water table, however, the opposite is true. Even minor heating of rock volumes

measured in terms of km^3 requires very large amounts of energy and, for natural circumstances, is conceivable only with the involvement of deep seated and hot or warm fluids.

Second, the veins from bore holes USW G-1, G-2 and G-3/GU-3 were found to contain two-phase (liquid and vapor) inclusions. Upon heating, these inclusions yielded homogenization temperatures ranging from 57 to as much as 245°C, Fig. 9 through 12 and Appendix E. Importantly, specimens representing the second isotopic facies were also found to contain fluid inclusions with shrinkage vapor bubbles. The homogenization temperatures of these inclusions fall in a range from 57°C to 227°C. The corresponding depositional temperatures are at least 30 -100°C higher than the contemporary ambient temperatures. This confirms the epithermal origin of the second isotopic facies, as inferred independently based on the observed $\delta^{18}\text{O}/\text{dz}$ gradient. At this point, an appropriate question may be asked: Are the homogenization temperatures reliably representing ambient temperatures at a time of either growth of host crystals (primary fluid inclusions) or subsequently to it (pseudo-secondary fluid inclusions reflecting epigenetic bathing of the crystals in fluids with temperatures and chemical compositions different from those of the parent fluids)? The following three points need to be considered in formulating an appropriate answer to this question:

- a homogenization temperature of primary fluid inclusions corresponds to a minimum possible temperature of a mineral forming fluid;
- a minimum temperature of entrapment of two-phase fluid inclusions can not be lower than 40°C, otherwise shrinkage bubbles will not form (this minimum

temperature is still 15°C higher than the corresponding contemporary ambient temperature); and

- in bore hole USW G-2, fluid inclusions that were tested for homogenization temperatures occur as pairs and yield comparable heating results; this implies the inclusions are neither stretched nor leaked and are recording paleotemperatures truthfully.

In light of these remarks, it is impossible for an objective interpreter to regard the fluid inclusion data as unreliable in at least confirming the independently inferred epithermal origin of the veins.

Third, both the epithermal origin of the second isotopic facies and the reliability of fluid inclusion data are further verified by ^{18}O calcite/opal geothermometry. As shown in Fig. 9 and 11, seven pairs of co-existing calcite and opal were tested for their ^{18}O signatures. These pairs were obtained from the vadose zone and paragenetically represent the second isotopic facies. The $\Delta_{\text{O-C}} = 0.74 (10^6/T^2) - 0.60$ relation (for water $\rightarrow$ minerals fractionation effects) gives depositional temperatures for the calcite/opal pairs in a range from 43°C (316°K) to 224°C (497°K). This range is in agreement with the 57-227°C range indicated by the homogenization temperatures of fluid inclusions. From Fig. 10 and 12, it may be observed that the vein set II appears to be developed as two thermally distinct facies TFa and TFb. For the TFa facies average depositional temperatures, obtained independently based on either fluid inclusion studies (79°C) or ^{18}O geothermometry (74°C), agree with each other. Similarly, for the TFb facies maximum depositional temperature of

227°C, indicated by the fluid inclusion data, agrees with corresponding temperature (224°C) given by ^{18}O geothermometry.

In summary, three independent and seemingly reliable lines of evidence lead to the same conclusion that is, the calcite/opal veins from the second isotopic facies were formed in the phreatic zone and in an epithermal environment. These veins mark the maximum elevation attained by the water table at Yucca Mountain during the Quaternary time. They must be considered in establishing the feasibility of the proposed high-level nuclear waste repository at Yucca Mountain.

Flooding of the Proposed Repository as an Anticipated Process

The paper by Quade has a direct bearing on an outstanding issue involving the likelihood of the occurrence of potentially catastrophic flooding of the proposed high-level nuclear waste repository. Specifically, in the context of long-term performance requirements set forth in 10 CFR 60, the paper provides geologic input into establishing whether or not flooding of the proposed repository constitutes an "anticipated process and event." By erroneously limiting "...the maximum elevation attained by water tables in the region during the Quaternary" to "...<115 m. and in most cases much less" and by implying that global climatic changes were solely responsible for the evident fluctuations of potentiometric levels, Quade leads a reader to believe that repository flooding should not be regarded as the anticipated process and event. Even putting aside the issues presented by the Whamonia mound and by the epithermal veins from the vadose zone, however, this conclusion

can not survive even a superficial scrutiny.

To illustrate this point a variety of information, obtained and presumably accepted by the Yucca Mountain scientific establishment, may be used. In this regard, a starting point is recognition that, for a period of 10^4 years, the likelihood of a combined volcanic intrusion and a fault rupture, occurring somewhere in the vicinity of Yucca Mountain, is better than one chance in ten. This estimate may be derived from results of paleoseismic investigations of faults on eastern Yucca Mountain (Menges et al., 1994 and Paces et al., 1994) and from results of radiometric dating of volcanics associated with the Lathrop Wells center (Wells et al, 1970; Turin and Champion, 1990). For the 10^4 year period, for which effective isolation of waste from biosphere must be assured, the probability of the return of a cool climate is near 100 percent. Thus, it is justified to regard a scenario whereby the combined volcanic intrusion-fault rupture events occur during a period when cool climate prevails, and to consider that this scenario has a probability of occurrence of about 10 percent. Accordingly, this scenario constitutes an "...anticipated process and event" and needs to be considered in evaluating compliance of the proposed repository with the regulatory requirements. The magnitude of the corresponding rise of the water table is a sum of three components: a) the rise of the water table caused by increased recharge during cooler climate; b) the rise of the water table resulting from a reduction of hydraulic conductivity around a fault ruptured nearby; and c) the rise of the water table resulting from invigorated or renewed thermal circulation of crustal fluids. As shown in Fig. 13, a rough estimate of the anticipated rise yields magnitudes ranging from 323 to 735 m, sufficiently large to flood the

proposed repository. As demonstrated in the attached Appendices F and G, magnitudes of the individual components were estimated using either observations made around Yucca Mountain (climatic change and fault rupture induced alteration of hydraulic conductivity) or results of calculations made by members of nuclear waste management community (thermal circulations).

Concluding Remarks

In summary, papers published by Quade (1994) and Quade and Cerling (1990) contribute to the dismissal of possible occurrences, at Yucca Mountain, of potentially catastrophic upwelling of geothermal fluids. Because the papers present demonstratively erroneously drawn conclusions, this dismissal is certainly not warranted. With reference to the commitment of the United States Government to the Yucca Mountain site, however, what is warranted at this time is a word of caution. This word of caution is justified by multiple lines of evidence and pertains to a hazard that can not be mitigated and reduced to prudent and acceptable levels. Such hazard must be adequately addressed prior to expanding countless millions of dollars on resolving much less germane issues. For this course of action to materialize, however, the word of caution must first originate from the scientific community and then be transmitted to and understood by the Government. In this regard, the contributions of Dr. Quade and his mentor Dr. Cerling are disappointing indeed.

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Figure 1. Location map - Crater Flat Spring deposits (sampling sites 24, 25, and 40). From Hill et al., 1993.

Figure 3a. Configuration of the paleo-potentiometric surface as proposed by Quade, 1994; absence of diatomaceous deposits between Rt.#95 and C-N State Line indicates that this proposition is in conflict with field observations.

Figure 3b. Configuration of the paleo-potentiometric surface not in conflict with field observations; to account for the absence of diatomaceous deposits, it is necessary to assume that, locally the paleo-potentiometric surface was inclined steeper than the contemporary potentiometric surface.

Explanation:

- - cobble encrustation from calcareous bedrock, isotopic data from Quade et al., 1988.
- - cobble encrustation from non-calcareous bedrock, isotopic data from Quade et al., 1988.
- ⊕ - Whamonie mound, isotopic data from Harmon and D.O.E. (1993).
- ▽ - Crater Flat Spring deposit, isotopic data from Hill and Schluter, 1994.

Note: Isotopic data from Harmon (1993) and Hill and Schluter (1994) are provided in Appendix E, Figures E-1 and E-2.

Figure 4. Comparison of isotopic character of carbon incorporated in surficial cobble encrustations with that incorporated in the spring deposits from Crater Flat and Whamonie.

Explanation:

- - cobble encrustation from calcareous bedrock, isotopic data from Quade et al., 1988.
- - cobble encrustation from non-calcareous bedrock, isotopic data from Quade et al., 1988.
- ⊕ - Whamonie mound, isotopic data from Harmon and D.O.E. (1993).
- ▽ - Crater Flat Spring deposit, isotopic data from Hill and Schluter, 1994.

Note: Isotopic data from Harmon (1993) and Hill and Schulter (1994) are provided in Appendix E, Figures E-1 and E-2.

Figure 5. Comparison of isotopic character of oxygen incorporated in surficial cobble encrustations with that incorporated in the spring deposits from Crater Flat and Whamonie.

Figure 6a. Isotopic characters of carbon and oxygen incorporated in modern soil carbonates and in the controversial Trench 14 deposits. From Quade and Cerling, 1990.

Figure 6b. $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ time series from core DH-11. From Coplan et al., 1994.

Note:

- a) the paleo-geothermal reconstructions shown are based on the depth-distribution of values of the $\delta^{18}\text{O}$ ratio, as observed in boreholes USW G-1, G-2, G-3, GU-3, G-4, and UE-25b #1;
- b) the actually measured values of the contemporary geothermal gradient range from 18 to no more than 24°C/km (USW G-1 - 18°C/km; G-2 - 24°C/km; G-3 - 22°C/km; G-4 - 24°C/km, and UE-25 b#1 - 20°C/km, Sass et al., 1987) ;
- c) a factor of ≥ 1.5 discrepancy, between the observed and the reconstructed values of the geothermal gradient, indicates that the Yucca Mountain veins were formed exclusively in association with geothermal processes.

Figure 7. Results of paleo-geothermal reconstructions, based on isotopic character of oxygen incorporated in the Yucca Mountain calcitic veins. From Whelan and Stuckless, 1992.

range of homogenization temperatures of two-phase fluid inclusions.

depositional temperatures of calcite-silica pairs, estimated employing ^{18}O geothermometry.

homogenization temperatures of two-phase fluid inclusions.

TF_a

thermal facies a.

TF_b

thermal facies b.

dT/dz

contemporary geothermal gradient.

$\delta^{13}\text{C}_{\text{PDB}}$ or $\delta^{18}\text{O}_{\text{SMOW}}$ ratio.

Figure 8. Explanation of symbols employed in constructing Figures 9 through 12.

Figure 9. Borehole USW G-1, homogenization temperature of fluid inclusions and isotopic characters of carbon and oxygen incorporated in calcitic veins from Yucca Mountain.

Fluid inclusion data are from Bish (1989) and D.O.E. (1993); isotopic data are from Whelan and Stuckless (1991) and D.O.E. (1993), and contemporary geothermal data are from Sass et al., (1987).

Figure 11. Borehole USW G-3/GU-3, homogenization temperature of fluid inclusions and isotopic characters of carbon and oxygen incorporated in calcitic veins from Yucca Mountain.

Fluid inclusion data are from Bish (1989) and D.O.E. (1993); isotopic data are from Whelan and Stuckless (1991) and D.O.E. (1993), and contemporary geothermal data are from Sass et al., (1987).

Figure 12. Borehole UE-25 A-5, depositional temperatures and isotopic characters of carbon and oxygen incorporated in calcitic veins from Yucca Mountain.

Depositional temperatures are from Dublyansky (1994), isotopic data are from Whelan and Stuckless (1991) and D.O.E. (1993), and contemporary geothermal data are from Sass et al., 1987.

$$\Delta H_A = \Delta H_C + H_{hy} + \Delta H_{EG}$$

where

ΔH_A - magnitude of anticipated rise of water table, at Yucca Mountain;

ΔH_C - rise of the water table resulting from climatic change;

H_{hy} - rise of the water table resulting from renewed volcanism and hydrothermal activity;

ΔH_{EG} - rise of the water table resulting from volumetric strain of 10^{-3} - 10^{-2} .

$\Delta H_C = 134$ m, based on observations at southern end of Crater Flat;

$H_{hy} = 102$ - 500 m, based on information provided in Appendix F;

$\Delta H_{EG} = 87$ - 91 m, based on information provided in Appendix G.

$$\underline{\Delta H_A = 323 - 725 \text{ m.}}$$

Note: At Yucca Mountain, the contemporary water table occurs at a depth ranging from 200 to about 400 m.

Figure 13. Rough estimate of the anticipated rise of water table at Yucca Mountain.

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APPENDIX A

APPENDIX A

**"Spring Deposits and Late Pleistocene Ground-Water Levels in
Southern Nevada"**

by Dr. J. Quade of the University of Arizona

SPRING DEPOSITS AND LATE PLEISTOCENE GROUND-WATER LEVELS IN SOUTHERN NEVADA

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ABSTRACT

Ground-water discharge deposits dating to the last glacial cycle and to several earlier cycles crop out in at least ten valleys in the southern Great Basin. The elevation and distribution of these deposits allow us to reconstruct the elevation of the water table during periods of wetter climate over much of the region, including the area around Yucca Mountain, site of the proposed high-level nuclear waste repository. Results from areas undisturbed by recent ground-water pumpage reveal that water levels have varied by <115 m, and in most cases much less, in the latter half of the Quaternary. The extent of ground-water discharge during older wet cycles is similar in scale to discharge during the last full-glacial period. This places most of the proposed repository horizon at least 85 to 285 m above the maximum levels attained by the water table under full-glacial climates. During the late-glacial period ($\sim 11,500$ to 8000 B. P.), a pulse of renewed discharge, perhaps corresponding to the Younger Dryas event well-documented elsewhere, produced organic-rich mats and flowing streams in many localities.

INTRODUCTION

The presence of paleospring deposits in southern Nevada has been recognized for some time.¹⁻⁶ These distinctive deposits are important because they mark the maximum elevation attained by water tables in the region

during the Quaternary. This information is crucial to establishing the feasibility of a repository for high-level nuclear waste at Yucca Mountain. Until very recently there has been little published evidence to support or refute predictions of small water-level changes in response to shorter-term climate change or other factors. During the last glacial cycle, moisture fluxes over the entire Great Basin were greater than today's. In response, water tables rose throughout the region. A pivotal but unresolved question for Yucca Mountain is how these water-level changes would effect repository integrity and radionuclide travel times in a future return to wetter, full-glacial conditions. The presence of fine-grained deposits in what are now dry valley bottoms in the southern Great Basin bears witness to spring discharge over large areas 17,000 yrs ago. Here we present new data on these deposits from 10 valleys (Fig. 1) that allow detailed reconstruction of water-table levels over the entire region during the last glacial maximum ("FG"; $\sim 17,000$ B. P.) and late-glacial period ("LG"; $\sim 11,500$ - 8000 B. P.).

STUDY METHODS

Spring deposits were mapped on aerial photographs ($\sim 1:60,000$), and this information transferred to topographic maps (1:24,000). The hydrologic environments were reconstructed using a combination of sedimentologic and faunal (mainly mollusks) evidence, as described in (6). We relied mainly upon carbonized wood and organic

Fig. 1 Major valleys and mountain ranges of the southern Great Basin. Dark triangles denote areas containing ground-water discharge deposits that we studied.

mats for ^{14}C dating of the deposits. All ages cited in this paper refer to ^{14}C yrs before 1950 A. D. Modern water table data was derived from several published sources (Table 1) and from unpublished well logs. A more complete account of our field and laboratory methods is presented in (7).

THE SPRING DEPOSITS

The spring-related sediments in all the valleys we studied share several characteristics that set them clearly apart from fluvial and lacustrine deposits already well characterized in the geological literature. Understanding these environments is crucial to reconstructing the elevation of former water tables. First, the fine-grained sediments in the southern basins don't fill the valleys like a bathtub ring. For example, the highest occurrences of these sediments mimic undulations in topography along the length of the Las Vegas Valley, but at levels generally 25 to 45 m above the current

water table. The dominant lithology, particularly just above the valley bottom, is cross-rippled silts containing dispersed cicada burrows (Fig. 2). These sediments lack mollusks except for a very rare *Vallonia cyclophorella*. The silts grade basinward into pale-green to white mudstones in the valley bottom. The mudstones usually contain a mixture of aquatic, semi-aquatic, and terrestrial mollusks. We examined 142 samples from 89 sites in ten valleys. Terrestrial mollusks such as *Vallonia cyclophorella*, *Pupilla muscorum*, and *Discus cronketei* are common. Typical semi-aquatic mollusks include *Gyraulus circumstriatus*, *Stagnicola caperata*, and *Succinea* sp. Only a few outcrops contain exclusively aquatic mollusks, such as *Planorbella subcrenata*, *Physa virgata*, *Valvata humeralis*, and *Gyraulus parvus*. These faunal patterns are to be expected in a system where moist ground and protective vegetation are abundant, but where standing water is of restricted extent.

Complete analogs to this mixed suite of paleontologic and sedimentologic features can be found in many basins of northeast Nevada.⁶ As with southern Nevada, transmissive Paleozoic limestones dominate the ranges bordering the deep basins, insuring that a high proportion of the moisture falling in the mountains reaches the basin through the subsurface, and not as surface streams. Valley cross-sections illustrate the common traits of the modern and fossil ground-water discharge systems (Fig. 2). Gravels cover the basin margins, and they grade abruptly into the fine-grained deposits on the distal portions of alluvial fans. In NE Nevada, the fine-grained deposits coincide broad, dense stands of non-obligate phreatophytes such as tall sage (*Artemisia tridentata*), rabbitbrush (*Chrysothamnus nauseosus*), and greasewood (*Sarcobatus vermiculatus*). The phreatophytes entrap fine sediment, preventing gravels from encroaching onto the valley bottoms. In the Steptoe and other modern valleys, the line defining this transition from coarse to fine-grained sediment essentially mimics the water table 6-7 m below the surface.⁸ As this textural transition is well exposed in full-glacial sediments in southern Nevada, we can use it to reconstruct the elevation of the top of water

Plant and Sediment Relationships

Fig. 2 The suite of sedimentologic features coinciding with different vegetation zones associated with a high water table. Sediments associated with phreatophyte flats are the most extensive of the fine-grained deposits in southern Nevada. The transition from coarse-to-fine sediment marks the position of the water table 6-7m below the surface in the Steptoe Valley of NE Nevada. This relationship can be used to reconstruct the paleowater table elevations in the southern basins.

TABLE 1 WATER-LEVEL CHANGES SINCE THE LAST FULL-GLACIAL PERIOD (FG)

AREA	WATER TABLE CHANGE (Δ WT in meters)	COMMENTS [^]	SOURCES OF WELL DATA*
Corn Creek Flat	25	FG marsh on lower flat	SE
Three Lakes Valley	30 to 40	FG phreatophyte flat, south end of valley	ref. 10; SE
Indian Springs Valley	60 to 70	FG phreatophyte flat, west end of valley	ref. 10; SE
Pahrump Valley	40	FG phreatophyte flat, upper Stump Springs drainage	ref. 18
Piute Valley	>30	minimum estimate due to dry well	ref. 19
S. Coyote Springs Valley	95	near Air Force Well #1	SE
N. Coyote Springs Valley	15	FG spring are now covered by mesquite	SE
Sandy Valley	10 to 20	FG spring are now covered by mesquite, north end	SE
Lathrop Wells Diatomite	<115m	maximum change since FG	ref. 10

* SE = water level data from well logs recorded at State Engineers office, Las Vegas, Nevada

[^] FG = full-glacial (~17,000 B. P.)

tables throughout the region 17,000 yrs ago. In the center of the basins in NE Nevada, white to green muds accumulate in wet meadows, springs, and in shallow marshes, lateral to the phreatophyte-covered flats described above (Fig. 2). They contain virtually the same mollusks today as found as fossils in green mudstones in southern Nevada. The mudstones also therefore serve as markers of formal water levels in the region.

A very detailed paleohydrologic reconstruction within single time periods is possible using these all these lines of evidence. As one example, fine-grained deposits are well exposed and cover about 25 km² on Corn Creek Flat north of Las Vegas. Nearly all the deposits date to the FG, except for a few outcrops at the south end that may date to an older pluvial cycle. During the FG, marshes covered about 8-12% of the 25 km² area, wet meadows about 21-25%, and dry flats vegetated by phreatophytes such as rabbitbrush and greasewood about 65%. Springs in this area also supported a through-flowing perennial creek, numerous smaller ponds and another small marsh around Tule Springs to the south.¹ Extensive discharge 17,000 B. P. also occurred in the Pahrump Valley, west of the Spring Mountains. Here at least two perennials creeks supported by spring discharge flowed on the southern end of the valley, although it is unlikely the creeks reached the now dry playa in the valley bottom. The LG spring deposits in the Pahrump Valley are inset into older springs deposits belonging to previous cycles of discharge extending at least into the middle Pleistocene.

Other valleys in the region containing spring discharge deposits include: Three Lakes Valley, Indian Springs Valley, Chicago Valley, Sandy (Mesquite) Valley, Coyote Springs Valley, Lathrop Wells, and Valley Wells (Fig. 1). While these deposits are not as abundant as those flanking the Spring Mountains, they provide important additional constraints of water level changes in the last 17,000 yrs.

WATER-LEVEL CHANGES AND THE IMPLICATIONS FOR YUCCA MOUNTAIN

Modern ground-water levels have been modified by pumping over the last century in some of the areas we studied. We therefore confined our comparisons of modern and paleo-groundwater levels to areas undisturbed in recent times. That virtually eliminated the lower Las Vegas Valley, the northern Pahrump Valley, and the central Sandy Valley, areas of extensive discharge 17,000 yrs ago, from our comparison. Paradoxically, our knowledge of full-glacial water levels is far more complete than that of modern water levels, given the wide dispersal of modern wells, and the questionable quality of some well records.

Our comparison reveals that ground-water levels have witnessed only modest changes (Δ WT) over the past ~17,000 yrs, from 5 to at most 115 m (Table 1), far short of the >500 m estimates of others.⁹ To cite a few examples, Δ WT has been about 25 m in the bottom of Corn Creek Flat, once covered by the FG marsh. In the southern Three Lakes Valley, the present water table is 30 to 40 m below FG pond sediments; it is 60 to 70 m below similar beds at the west end of the Indian Springs Valley. The focus of our comparison is on the highest elevation occurrences of the fine-grained deposits along basin margins. This is because Δ WT can to increase with distance away from the basin center. The Stump Springs drainage in the southern Pahrump Valley illustrates this point. During FG and LG times, a perennial creek, fed largely by paleosprings located upslope, flowed over the current location of the springs. So our estimate of Δ WT at this point would be close to 0 m. Δ WT is about 40 m where fine-grained deposits first appear several kilometers upgradient. In these ground-water discharge systems, creeks, wet meadows, marshes, and even some springs often lie well below the level of intersection of the *top* of the regional water-table with the land surface. Comparisons of changes below that level of intersection therefore should represent minimum estimates of Δ WT.

The closest unequivocal spring deposits to Yucca Mountain are what we informally

Fig. 3 The Lathrop Wells diatomites are the closest unequivocal ground-water discharge deposits to Yucca Mountain yet identified. The position of the diatomites requires a ≤ 115 m rise in the water table during the Quaternary. These relationships can be extrapolated to Yucca Mountain if one assumes that water table gradients did not change substantially in moister climates. If valid, then the proposed repository horizon remained 85-285 m above the regional water table throughout much of the Quaternary.

designate the Lathrop Wells diatomites. They are composed of a least three outcrops of locally diatomaceous, tuffaceous sediment lying at the southern margin of Crater Flat, about 20 km SW of the proposed location of the waste site at Yucca Mountain. We regard these sediments as originating from discharge from the regional water table directly out of local volcanic rocks, and therefore they are key indicators of water levels in the vicinity of Yucca Mountain during the Quaternary. The age of the Lathrop Wells diatomites is still under discussion. At this point, there is no compelling field evidence for ground-water discharge at this location during the last glacial cycle, although the paleontological evidence for a Pleistocene age is firm. The extent of the deposits and nature of the fauna in the Lathrop Wells diatomite indicates that regional saturation was at this level for some periods in the pre-Wisconsin. The present water table is about 115 m below the highest of the Lathrop Wells deposits on southern Crater Flat,¹⁰ implying the FG rise of the water table was less than that in this area. Our estimate for $\Delta W T$ at this locality during the last glacial cycle is < 115 m (Fig. 3). This estimate can be extrapolated to the proposed repository site 20

km away, provided water table gradients have not changed significantly. The present gradient on the water table between the site and the Lathrop Wells diatomites is ~ 1 m/km. This modest gradient is a result of (1) high hydraulic conductivities of the fractured welded tuffs in the aquifer, (2) subdued topographic relief of the area, (3) the large distance to potential recharge areas north of Yucca Mountain. These considerations make it likely that increased flow cross-sections, and not increases in water-table gradients would be the principal factor responding to enhanced recharge. If valid, this indicates that most of the repository remained at least 85 to 285 m above the level of the regional water table during the last glacial cycle, and probably as far back in the Quaternary as the age of the Lathrop Wells diatomites turn out to be. This estimate agrees closely with most isotopic and mineralogical evidence for modest $\Delta W T$ from boreholes within Yucca Mountain (summarized in (7)).

It is more difficult to predict the effects of enhanced recharge on the water levels under the northwest portion of the repository, where the regional water table gradient steepens substantially. The exact cause for the gradient

change is unknown but may be related to low transmissivity along a deeply buried fault.¹¹ The steepened gradient across the low transmissivity zone could result in a larger ΔW (>115 m) in a return to glacial climate than elsewhere under the proposed repository horizon.

LATE GLACIAL SPRING DISCHARGE: A RAPID AQUIFER RESPONSE TO THE YOUNGER DRYAS RECHARGE ?

One of the most conspicuous features of deposits belonging to the latest glacial period are organic-rich "black mats". We have described and dated these features in five of the valleys in the region, and they have been also described in the lower Las Vegas Valley.¹ The black mats consist of 1-5% (by weight) C mixed with silt and fine sand. They vary in thickness, from a few centimeters up to a meter, and in nearly all cases lack internal bedding. The mats can be continuous laterally for 10's of meters, or more localized and cauldron shaped. Mollusks, both terrestrial and aquatic, are common but not necessarily always present. The mollusks found in the mats are consistent with moist terrestrial to marshy conditions typical of areas around spring discharge points and their outflow channels

More than twenty ^{14}C dates on the humate fraction of organic matter range from 11,800 to 2000 B. P. The majority of ages on the mats cluster between 11,000 and 10,000 ^{14}C yrs B. P. (Fig. 4) The black mats typically overlie green clays probably dating to the full-glacial, and are overlain by brownish silts of early Holocene age. A weakly developed soil often separates the mat from underlying full-glacial marsh deposits. This succession suggests that: (1) drying of some areas preceded the renewal of vigorous spring activity just before 11,500 ^{14}C yrs B. P., (2) renewed spring discharge was greatest between 11,000 and 10,000 yrs B. P., (3) discharge continued but tapered off gradually between 10,000 and about 7200 B. P., and (4) complete drying of most spring areas was complete no later than 7200 B. P.

The abrupt appearance of black mats in the record coincides with other indicators of

Fig. 4 Histogram of ^{14}C ages (in years B. P.) on the humate fraction in black mats from southern Nevada.

renewed discharge starting shortly before 11,000 ^{14}C yrs B. P. At Tule Springs new channels were cut and perennial water was flowing starting about 11,200 B. P. The same had occurred on lower Corn Creek Flat shortly before $10,980 \pm 270$.⁴ In the Pahrump Valley, spring-fed channel flow recommenced just prior to 11,200 in one and perhaps two major stream courses,⁷ following an apparent hiatus in perennial flow of unknown duration.

The renewal of discharge associated with the black mats shortly before 11,000 B. P. correlates closely with the Younger Dryas (YD) cooling event, as first pointed out by (12). The YD event, spanning the period ~11,000 to 10,000 yrs B. P.,¹³ was first recognized in northern Europe but is now well-documented in many areas of the world, and with exceptional detail recently in the Greenland ice cores.¹⁴ It is also now described from oxygen isotopic evidence in marine cores from the Gulf of California.¹⁵ The close correlation between the known ages of the YD event and the reinitiation of vigorous spring flow in the southern Nevada basins strongly suggests that the aquifers feeding the

valley discharge systems were capable of rapid response (10's to a few hundreds of years) to the changes in recharge. The YD event is thought to be related to processes directly related to deglaciation.^{16,17} However, rapid changes (or "flickering" as it has been recently described) in interglacial climate are now documented to occur on a scale of tens of years.¹⁴ Future Holocene climate may be far less stable than the last 8000 yrs indicates. If these changes are of sufficient magnitude (i. e. like the YD event), our evidence suggests that quick response at discharge points is likely. At the same time, we observe that even the YD event, while conspicuous in the southern Nevada record, led to water table increases no higher than that of the last full-glacial period, and then in only restricted areas.

CONCLUSIONS

Effective moisture was greater during the full and late-glacial periods in the southern Great Basin, and in response water tables rose throughout the region. The extent of that rise varied from location to location but 115 m of change is an upper limit, and less than 100 m is typical. This places the repository well above the water table in the event of a return to glacial climate, if water level changes observed in the basins can be validly extrapolated into the volcanic bedrock terrane at Yucca Mountain. This is a key uncertainty that should be borne in mind by future studies. Our present analysis would benefit in particular with more water-level information in the area between the Lathrop Wells diatomites and Yucca Mountain, as well as directly under the diatomite itself. Our analysis also urges careful consideration of the causes of the low transmissivity zone underlying the NW portion of the repository and the possible consequences for Δ WT with a return to glacial climate.

The strong correlation between the age of the black mats and the Younger Dryas Event suggest that the valley aquifers in the region are capable of rapid response to changing recharge conditions.

The intense focus over the last several years on the question of changes in the

regional water level in response to climate change at Yucca Mountain has deflected attention from other unresolved paleohydrological issues, such as the effects of a marked increase in fracture flow and extent of perched ground water that will attend a return to glacial climate. This key area has not been effectively addressed thus far, and should remain a focus of ongoing paleohydrological research at Yucca Mountain.

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APPENDIX B

APPENDIX B

Summary, Conclusions, and Recommendations from the 1989 report by J. S. Szymanski

Conceptual Considerations of the Yucca Mountain
Groundwater System with Special Emphasis on
the Adequacy of This System to Accommodate
A High-Level Nuclear Waste Repository

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SECTION 5.0

SUMMARY, CONCLUSIONS, AND RECOMMENDATIONS

5.1 SUMMARY AND CONCLUSIONS

Having performed the analyses presented in Sections 3.0 and 4.0, the following three main conclusions seem to be reasonable and warranted at this time:

i) Examination of the extensive data base pertaining to the Yucca Mountain groundwater system, in light of the conceptual framework as established in Section 3.0, reveals that this system is considerably different from the flow system currently envisaged by the Yucca Mountain Project. The conceptual understanding of the groundwater system, as used in performing suitability assessments for the purposes of developing the Final Environmental Assessment (DOE, 1986) and in establishing an approach to the forthcoming site characterization, is far too simple and much too far removed from reality. Simply stated, this conceptual understanding completely ignores the volcano-tectonic setting of the Yucca Mountain site.

According to the currently accepted conceptual understanding, the Yucca Mountain groundwater system is a "simple" system that may be represented by one of the commonly used mathematical models of groundwater flow (Section 2.2). The water table is regarded as a meaningful and durable expression of the hydraulic conductivity structure and boundary conditions. Position of the water table is assumed to be related solely to the volume of fluid passing through the system's vertical boundaries. Such a volume of fluid is a function of climatic conditions and remains fixed as long as these conditions remain unchanged. It is assumed that all moisture in the vadose zone is the result of infiltration of a small portion of local rainfall. Furthermore, it is assumed that movement of water in the vadose zone involves matrix flow, and that such movement occurs under the influence of gravity alone.

Within the context of currently accepted conceptual understanding, the influence of tectonic disruptions is limited to a) local alteration of litho-stratigraphic relationships; b) additional local fracturing; and c) minor changes in altitudes of discharge areas relative to altitudes of recharge areas. Magnitudes of tectonically induced changes in these factors are relatively small; consequently, the resulting alterations of the hydraulic conductivity structure and the boundary conditions are likewise small. Such alterations may

readily be accommodated by minor changes in position and configuration of the water table. It follows then, that the impact of such alterations on the long-term performance of a high-level nuclear waste repository is negligible.

A completely different picture emerges if one considers the volcano-tectonic setting of the Yucca Mountain Site and how it affects the local groundwater system. As indicated in Section 3.4.1, the area of Yucca Mountain contains continental alkalic basalts of Plio-Quaternary age. These basalts, as a matter of fact, are trace-element enriched hawaiites with isotopic compositions indicative of upper mantle origin. The isotopic and geochemical characteristics of the basalts are best accounted for by assuming that a convective system is developed in the upper mantle of this region. The presence of such a system may be inferred independently based on interpretations of various geophysical data (Sections 3.1 and 3.4.1). From the perspective of local and regional hydrology, the upper mantle convective system is responsible for two important tectono-physical factors. The first factor is high, and most importantly, spatially heterogeneous flux of terrestrial heat. Terrestrial heat and gravitational hydraulic pressures acting at vertical boundaries of the groundwater system are the energy sources, that together, are responsible for the movement of groundwater. In the thermodynamic sense, groundwater flow is a heat-fluid coupled flow. For a groundwater system involving such flow, any plane adopted as the base must be considered as a "variable-upward-flux" boundary with respect to both heat and fluid. Mathematical models used in numerical simulations of such a flow system must account for this circumstance. Otherwise, calculations and interpretations may be grossly misleading, and with reference to the Yucca Mountain groundwater system, are quite irrelevant.

The second factor is strain energy, that because of the upper mantle convective system, is being supplied into the groundwater system on a continuous basis. The resulting changing *in situ* stress field plays an important role in controlling the resistance of a fractured medium against the coupled flow of heat and fluid. Near the ground surface, the *in situ* stress field controls the conducting apertures of fractures, and through that, it also controls the local effective thermal and hydraulic conductivity structures. For time periods of geologic proportions, the entire groundwater system behaves as a three way, H-F-M, coupled thermodynamic substance. For purposes of these conceptual considerations, this substance has been simplified and referred to as a heat-fluid coupled flow system in a deforming fractured medium. The substance evolves through a series of similar evolutionary loops, that for a given portion of the groundwater system, are closely tied to cycles of tectonic deformation (Section 3.3.2). During a single evolutionary

loop, energy content of the substance undergoes systematic transformations (Section 3.5.3). At the onset of a deformation cycle, the *in situ* stress gradients are fairly low with few aberrations. Such a state of *in situ* stress allows conducting apertures of fractures to be low. The resulting low hydraulic transmissivity and low effective thermal conductivity allow for the existence of high gradients of hydraulic potentials and *in situ* temperature. At that time, the water table is high and the flow system exhibits strong "mixed" convection aspects. As deformation progresses, "steeper" *in situ* stress gradients begin to emerge. In some areas around yielding fractures, the stress difference diminishes progressively and eventually produces "isotropic" and "singular" points. Conducting apertures of fractures increase in a time-dependent manner (Section 3.3.3.3). Such time-dependent increases cause the gradients of temperature and hydraulic potentials to "weaken." The water table deepens, and near the ground surface, the groundwater system loses its strong "mixed" convection or geothermal aspects. The termination of a deformational cycle is associated with a significant rearrangement of the local *in situ* stress field. The rearranged and "weakened" *in situ* stress field allows for reappearance of the normal "strong" gradients of *in situ* temperature and hydraulic potentials. The water table returns to its former position and the convective flow of heat and fluids returns to its normal intensity. At that time, the bedrock separations that accompanied the *in situ* stress field rearrangements are being filled with minerals precipitating from convecting fluids.

As demonstrated in Section 4.0, the assumption of a heat-fluid coupled system in a deforming fractured medium is in full accord with all known characteristics of the Yucca Mountain groundwater system. This is in sharp contrast to the conceptual understanding of this system as used to a) perform assessments of the suitability of Yucca Mountain for site characterization and b) develop strategies for demonstrating compliance of this site with all applicable Federal requirements. Not accounting for the true nature of the local groundwater system resulted in serious errors in judgment regarding anticipated long-term performance of the proposed high-level nuclear waste repository.

ii) **Conceptualization of hydrologic processes operating in the Yucca Mountain vadose zone yields a completely different picture than the one currently envisaged by the Yucca Mountain Project.** Two issues of fundamental importance are associated with the possibility that the Yucca Mountain groundwater system is developed in a deforming fractured medium (i.e., the mechanism of flow in the vadose zone and temporal stability of the water table, including its short- and long-term aspects, and involving both climatic and tectonic factors). Two other issues of nearly equal importance are associated with the possibility that the Yucca Mountain groundwater system involves a heat-fluid coupled

flow (i.e., chemistry of interstitial water in the vadose zone and spatial and temporal variability of flux of terrestrial heat through the vadose zone).

A relative measure of likelihood of fracture flow in the vadose zone is depth extent of the "limit equilibrium" stress conditions (Section 3.3.5). The results of *in situ* stress determinations, made at Yucca Mountain, revealed that the depth extent of such conditions is considerable, in a range of 1,200 to 1,500 m (Section 4.2.5.4). This indicates, that in the Yucca Mountain vadose zone, conducting apertures of fractures are very large. It is likely that such apertures may support local fracture flow, despite a low annual average rate of infiltration of rainfall. Furthermore, in the Yucca Mountain vadose zone, the state of *in situ* stress is such that numerous "singular" points are present. Around such points the mean stress and the stress difference are very small and, consequently, any build-up of fluid pressure causes further enlargement of local conducting apertures. It is unlikely that stressed in such a manner fractured medium may retain fluids for a period of time that is required for introduction of these fluids into rock matrix. Both of the above lines of reasoning suggest that the expectation of solely interstitial flow in the Yucca Mountain vadose zone lacks proper foundation. This is true, in particular, if one is concerned with movements of water through rocks with either high saturation or low interstitial porosity, as is the case for the tuffaceous beds of Calico Hills and for the densely welded Topopah Spring Member of the Paintbrush Tuff, respectively.

In a deforming fractured medium, temporal stability of the water table is not related solely to global or regional climatic fluctuations. Such stability is also related to tectonic factors, specifically the local *in situ* stress field. As discussed in Section 3.3.5, both short-term and long-term instabilities of the water table may be expected at the Yucca Mountain site. The short-term instabilities involve several meters, perhaps tens of meters, of displacement of the water table. They are short-lived, say days or weeks at most, and occur with a frequency related to a degree of development of the local *in situ* stress field. Such short-lived instabilities are caused by continuous but minor restructurings of the local *in situ* stress field (Sections 3.3.3.3, 3.3.4, 3.3.5, and 3.3.6). They may occur spontaneously, or they may require an external triggering mechanism. Fault creep, vibratory ground motion induced slip, local and temporary volumetric strain, are examples of minor restructurings of the local stress field. As discussed in Sections 4.1 and 4.2.4.4, the results of continuous monitoring of the water table, performed at Yucca Mountain in a number of deep wells during the last six years, are not yet available in an accessible and reliable form. Consequently, it is not known with certainty whether or not the short-lived instabilities of the water table are

currently occurring at the Yucca Mountain site. Such instabilities, however, are of major diagnostic importance. They may confirm directly that the Yucca Mountain groundwater system, indeed, operates in a deforming fractured medium. Also, the short-lived instabilities, if occurring with a meaningful frequency and magnitude, constitute a "pumping" mechanism for gaseous transport through the vadose zone.

The long-term displacements of the water table are caused by large-scale restructurings of the local *in situ* stress field. Such restructurings occur at the end of tectonic cycles and, at Yucca Mountain, with a periodicity measured in terms of tens of thousands of years. A magnitude of the resulting potential displacements of the water table is a sum of three components: a) rise of the water table caused by "overpressure" Δp_{max} (Section 3.3.4 and Plate 3.3.5-1); b) rise of the water table resulting from reduction of storativity (Section 3.3.5 and Plate 3.3.5-1); and c) rise of the water table resulting from tectonically induced restructuring of the local hydrothermal system (Sections 3.3.4 and 4.3.3; Plate 3.4.4-6). The existing hydrologic and geothermal data indicate

- 1) The greatest magnitude of "overpressure," that is known to exist at Yucca Mountain, is $\Delta p = 55$ m (Plates 4.2.4.4-12 and 4.2.4.4-17a). The total magnitude of Δp_{max} , however, is unknown. It is possible, nevertheless, that this magnitude is approximately equal to the difference in altitudes of the water table in northern and southern sectors of Yucca Mountain. As shown on Plate 4.2.3-5, this difference is equal to at least 300 m;
- 2) At Yucca Mountain, potential build-ups of fluid pressure that are caused by fluids released from storage may be very large, and measured in terms of tens of bars. Such a state of affairs is suggested by two lines of evidence. First, the ratio S^o/S^{in} appears to be surprisingly large, within a range of 10^2 to 10^4 (Section 4.2.6 and Plate 3.3.5-1). Second, below the zone of *in situ* stress enhancement of the local hydraulic conductivity structure, the average value of the lateral hydraulic gradient may be within a range of 10 to 20 m · km⁻¹ (Plate 4.2.3-2). In this situation, even for a relatively small zone of the restructured *in situ* stress field, the value of fluid pressure build-up that is required to "bleed-off" the volume of fluid released from storage is very large indeed; and
- 3) Potential rise of the water table resulting from a restructuring of the deep Yucca Mountain hydrothermal system may be significant and measured in terms of, at least, tens of meters. The following two lines of reasoning are offered in support of this statement. First, the geochemical and

isotopic characteristics of fluids encountered around the Solitario Canyon fault, specifically the $\delta^{13}\text{C}$ ratios and the calcite saturation index, may be interpreted as indicative of local upwelling of fluids from the Paleozoic carbonates (Section 4.3.3). Assuming that such upwelling is the result of thermally driven convection, it follows that the depth extent of the local hydrothermal system is equal to at least a local depth of occurrence of the Paleozoic carbonates. Based on interpretation of the results of magnetic and gravity surveys, depth of occurrence of the Paleozoic carbonates was estimated to be ranging from 3.4 km, at the northern end of Yucca Mountain, to 4.0 km in Crater Flat (Carr et al., 1986). Second, the results of isotopic analyses of vein mineralization from the Yucca Mountain vadose zone suggest that contemporary values of the vertical geothermal gradients are some $15^\circ\text{C}/\text{km}$ lower than those in the not too distant past (Section 4.3.4). Assuming that the discrepancy is, at least partially, related to past restructurings of the Yucca Mountain hydrothermal system then it follows, that the expected average change in water temperature may be in a range of few tens of degrees Celsius, say 20 to 30°C . In this situation, the potential "bouyant" rise of the water table alone is in a range of 32 to 48 m (Section 3.4.4).

The above crude estimates indicate, that at Yucca Mountain, magnitudes of the long-term displacements of the water table involve several tens and perhaps a few hundred meters. Displacements of the water table of such magnitudes are likely to profoundly impact the long-term performance of the proposed high-level nuclear waste repository. Also, in extreme cases such displacements may result in flooding of the repository and expulsion of groundwater at the land surface. A critical point in the overall conceptual considerations of the Yucca Mountain groundwater system is to establish unequivocally whether or not the geologic record representing Plio-Quaternary time contains expressions of the past large-scale displacements of the water table. As discussed in Section 4.5.3, there are indications that this may be the case indeed. Thus far, however, only superficial studies of the calcite-opaline silica vein deposits were performed during the last 10 years. Consequently, the current data base does not allow for an unequivocal resolution of this critical issue. Nevertheless, the possibility that the veins constitute a record of large-scale displacements of the water table in the past is real and, by all means, not remote.

The assumption that the Yucca Mountain groundwater system is developed in a deforming fractured medium and, that it involves a heat-fluid coupled flow, leads to another important issue: the origin and chemistry of interstitial waters in the Yucca Mountain vadose zone. If interpretations performed in this report are correct then it follows that interstitial waters represent, at least in part, relict hydraulic

mounds developed in association with high groundwater stands at the end of tectonic cycles. Chemistry of such waters should be different from that of "ordinary" meteoric waters infiltrating through the vadose zone. At Yucca Mountain, downward movement of infiltrating meteoric waters may involve interstitial flow, but only during early stages of deformation, when intensity of the local *in situ* stress field is low and conducting apertures of fractures consist of only residual components. Because of the high water level stand, however, it is unlikely that significant replacements of interstitial water, by infiltrating meteoric water, could occur at that time. Such replacements may be significant, but only during more advanced stages of deformation, when the water table is deeper. At that time, however, the intensity of *in situ* stress field is substantially increased and, consequently, conducting apertures are enlarged, so that infiltration of meteoric water occurs as fracture flow. As far as it could be determined, there are no published data concerning chemistry of interstitial water from the Yucca Mountain vadose zone. Consequently, there is no basis to either confirm or reject the hypothesis that interstitial waters are relict waters that originated from below the water table. Nevertheless, there are some unpublished data that seem to indicate that chemistry of interstitial waters is different from that of waters residing in fractures. Such data were obtained by Dr. A. Yang (USGS, Denver, CO) through chemical analyses of samples of interstitial waters that were mechanically squeezed out of cores extracted from the Yucca Mountain vadose zone.

The conceptual understanding of the Yucca Mountain groundwater system, as developed in this report, indicates that **flux of terrestrial heat through the vadose zone** lacks spatial uniformity and temporal consistency. Similar conclusions may be drawn based on the Yucca Mountain geothermal data. The downhole temperature profiles indicate that the local geothermal field contains very substantial lateral geothermal gradients (Section 4.3.3). The results of isotopic analyses of vein mineralization from the vadose zone suggests that potentials of the local geothermal field may be time-dependent (Section 4.3.4). Both of the above variabilities are caused by tectonic factors that control the local effective thermal conductivity structure and the local thermal boundary conditions. Clearly, the assumption of "const. flux" thermal boundary conditions, for the base of the vadose zone, cannot be justified. Moreover, the local intensity of flux of terrestrial heat is substantial, and may be within a range of 70 to 100 mWm⁻¹. In this situation, terrestrial heat is a factor that needs to be reckoned with in considering movements of fluids, vapors, and gases in the vadose zone.

iii) The conceptual understanding of the Yucca Mountain groundwater system, as developed in this report, points out serious limitations of the Yucca Mountain site to effectively

isolate radionuclides from the biosphere. Such limitations are greater by far than those currently envisaged by the Yucca Mountain Project. Without recognizing these limitations, the issue resolution strategies and performance allocations, as expressed in the Statutory Site Characterization Plan, are interesting propositions, but in terms of the actual site conditions are of limited practical value. The existing data indicate that it is very probable that full implementation of the proposed study plans will not yield data that are required for a successful completion of the licensing process. On the contrary, implementation of the proposed study plans will undoubtedly yield data that will bring limitations of the Yucca Mountain site into a full light. At that time, substantial changes in the site characterization "philosophy" will be required. Furthermore, not relying on timely and reasonable interpretations of the current data base, the Yucca Mountain Project failed to develop a realistic comprehension of the local hydrologic conditions. The resulting misunderstanding of hydrology of the Yucca Mountain vadose zone resulted in overly optimistic assessments of the suitability and licenseability of the Yucca Mountain site.

The most important licensing concern is a potential rise of the water table. If such a tectonically induced rise is possible at Yucca Mountain, then the geochronologic data indicate that the annual probability of occurrence of this rise is more than 10^{-5} , perhaps within a range 3 to 5×10^{-4} (Sections 4.3.4 and 4.5.3). It is likely, that in the context of long-term performance requirements as set forth in 10 CFR 60 and 40 CFR 191, the water table rise constitutes an "anticipated process and event." As such, the water table rise must be accounted for in demonstrating compliance of the Yucca Mountain disposal system with three long-term performance objectives of 10 CFR 60. Such performance objectives are a) the life of the waste package; b) the release rates of radionuclides from the "engineered barrier system"; and c) the overall releases of radionuclides into the "accessible environment." The water table rise has a potential of significantly altering radionuclide migration path, rate, and time. An occurrence of the water table rise during early stages in the life of the repository, when temperature of waste packages and the surrounding host rock is above the boiling point of water, entails a particularly profound impact on the overall repository performance. Vaporization of fluids entering the repository would cause substantial increases in the vapor pressure and, therefore, would accelerate gaseous transport from the repository to the ground surface. Subsequent cooling of the repository, to below the boiling point of water, is likely to be accompanied by a long-term convective flow of fluids from the repository to the ground surface.

It is expected, that chemistry of fluids entering the repository in association with the tectonic rise of the water table, is different from that of fluids currently residing in fractures. The impact of tectonic rise

on the life of waste packages, therefore, is two-fold. Corrosion of waste packages will be accelerated due to a) increased amounts of water contacting the packages and b) more adverse chemistry of this water.

A tectonically induced rise of the water table, with magnitudes sufficient to flood the repository, is not the only licensing concern that results from the proposed understanding of the Yucca Mountain groundwater system. This understanding brings into sharp focus serious reservations with regard to compliance of the Yucca Mountain site with the groundwater travel-time requirement as set forth in 10 CFR 60. Compliance of the Yucca Mountain site with that requirement is assured, but only if meteoric water infiltrates through the vadose zone via mechanism of interstitial flow. In the case of fracture flow, if infiltration occurs at all, the travel-time through the vadose zone is, most certainly, substantially less than the required 10^3 years.

A gaseous transport through the vadose zone is another licensing concern resulting from the proposed understanding of the Yucca Mountain groundwater system. Such a transport is caused by the short-lived instabilities of the water table and by the heterogeneous flux of terrestrial heat into the vadose zone. The gaseous transport may have some impact on the compliance of the Yucca Mountain site with two performance objectives of 10 CFR 60. Such performance objectives are a) the release rate of radionuclides from the "engineered barrier system" and b) the overall releases of radionuclides into the "accessible environment."

Finally, the proposed conceptual understanding of the Yucca Mountain groundwater system suggests that chemistry of interstitial waters in the vadose zone may be substantially different from that of fracture waters. Such differences may be important with regard to considerations of regulatory compliance of the Yucca Mountain site. Three long-term performance objectives of 10 CFR 60 may be impacted. Such performance objectives are a) the life of waste packages; b) the release rates of radionuclides from the "engineered barrier system"; and c) the overall radionuclide releases into the "accessible environment."

Most of the above licensing concerns are not new concerns. They were raised previously, in one form or another by various parties, most notably by the U.S. Nuclear Regulatory Commission and the State of Nevada. The proposed conceptual understanding of the Yucca Mountain groundwater system reinforces those concerns and provides a uniform theoretical background for them.

5.2 RECOMMENDATIONS

The proposed conceptual understanding of the Yucca Mountain groundwater system poses a number of serious questions regarding the suitability of the Yucca Mountain site to safely accommodate a high-level nuclear waste repository. Recognizing the importance of such questions, and keeping in mind the long-established policy of the DOE to always address the critical suitability issues first, the following eight recommendations are offered:

- 1) A thorough examination and reasonable interpretation of the data collected during the last six years of continuous monitoring of the Yucca Mountain water table should be performed without any further delays. A main purpose of this task is to establish unequivocally whether or not a) short-lived instabilities of the water table are present and b) response of the local water table to externally generated vibratory ground motions is abnormal. Furthermore, it is important to determine a frequency of occurrence of the short-lived instabilities as well as their magnitude;
- 2) A comprehensive investigation of the state of Yucca Mountain *in situ* stress field should be undertaken. Such an investigation may be performed utilizing the presently existing network of boreholes, and employing both the hydraulic fracturing technique and the Cooper-Brederhoeft injection tests. A main purpose of this task is three-fold: a) reliable and unequivocal detection of "singular" points; b) confirmation that the *in situ* values of closure pressure are abnormally low; and c) fine definition of spatial distribution of the *in situ* values of closure pressure. The measurements should be made using small values of the applied hydraulic pressure and employing various diameters of injection tubing;
- 3) A deep exploratory borehole should be drilled in the western sector of Yucca Mountain. The main purpose of this borehole is to gain further understanding of the depthward distribution of hydraulic potentials in this area. Specifically, it is of critical importance to determine a magnitude of the Yucca Mountain "overpressure." The borehole should be drilled to a depth that is sufficient to penetrate the Paleozoic carbonates for a distance of a few hundred meters. Below a depth of about 1,500 m, a series of drill stem tests should be made. After completion of the borehole, a few sets of packers and piezometers should be installed, so that the downhole measurements of hydraulic potentials could be made at discreet depth intervals;

- 4) A complete and conclusive investigation of the calcite-opaline silica-sepiolite vein deposits from Yucca Mountain should be undertaken without any further delays. A purpose of this task is to establish unequivocally whether or not the geologic record contains "foot prints" of past large-scale fluctuations of the water table. Specifically, a firm position must be established whether or not flooding of the repository, within the context of 10 CFR 60 and 40 CFR 191, constitutes an "anticipated process and event." The investigation should be carried out in both the surface and subsurface, and should include radiometric age determinations;
- 5) A thorough study of the Yucca Mountain "mosaic" breccias should be undertaken without further delays. A main purpose of this task is to establish whether or not a) the "mosaic" breccias record a substantial volumetric strain of the Yucca Mountain area, in a range of several percent and b) the "mosaic" breccias are younger than solidification of the country rock, polygenetic, and representing time spans measured in terms of 10^6 years. Specifically, a firm position must be established whether "widening" of Yucca Mountain is an ongoing and cyclic process. The studies should include limited trenching, high resolution mapping in selected areas, and petrographic analyses of selected samples to determine petrogenesis involved in formation of the "mosaic" breccias;
- 6) An investigation of chemistry of fluids residing in the Yucca Mountain vadose zone should be performed. A main purpose of this task is to establish whether local interstitial waters are chemically and isotopically different from fracture waters. The investigation should consist of chemical and isotopic analyses of samples of interstitial waters that were mechanically squeezed out from cores following the procedure used by Dr. A. Yang of the USGS. The core samples will be extracted from a few boreholes that should be drilled for that purpose;
- 7) An investigation of the known bodies of perched and semi-perched water, that occur within the vadose zone of the Death Valley groundwater system, should be performed. A main purpose of this task is to evaluate the possibility that some of these bodies represent tectonically induced hydraulic mounds, either remnant or active. There are three outstanding candidates for this study: a) the apparent hydraulic mound underneath Greenwater Range; b) the hydrologic anomaly in the vicinity of

Skull Mountain; and c) the apparent hydraulic mound between Climax Stock and Emigrant Valley.

The investigation should be sufficiently broad in scope to permit a definition of lateral and vertical gradients of a) hydraulic potentials; b) *in situ* temperature; and c) isotopic and chemical composition of groundwater; and

- 8) An investigation of the Whamonie gypsum paleo-spring mound should be performed. A main purpose of this task is to establish age of activity of the Whamonie paleo-spring discharge. Specifically, a firm position must be established whether or not the apparent upwelling of deep, sulfate rich, waters is of Late-Quaternary age. The investigation should include high resolution mapping, limited trenching, petrographic analyses of selected samples, and radiometric age determinations.

The recommended investigations were designed to obtain data that are judged to be required for validating the proposed herein conceptual understanding of the Yucca Mountain groundwater system. The presence of a) short-lived instabilities of the water table and abnormal responses of the water table to externally generated vibratory ground motions; b) "singular" points and abnormally low values of the closure pressure; c) sizable "overpressure"; d) "foot prints" of past large-scale fluctuations of the water table during Quaternary time; e) post-depositional and polygenetic "mosaic" breccias; f) meaningful differences in chemical compositions of interstitial and fracture waters; g) perched and semi-perched waters resulting from injection of fluids from below the water table; and h) ground surface discharges of warm, sulfate rich, fluids during Late-Quaternary time, would confirm that the proposed understanding of the Yucca Mountain groundwater system is correct. In this situation, and within the context of current Federal regulations, serious considerations should be given to abandoning the Yucca Mountain site and declaring it as unlicenseable for the purposes of permanent disposal of high-level nuclear waste. At a minimum, a fundamental revision of the strategies for demonstrating regulatory compliance of the Yucca Mountain site, as developed in the Statutory Site Characterization Plan, will be required. In all sincerity, the DOE would be well advised to perform the recommended investigations in a truly expeditious manner, preferably prior to the commitment of substantial resources such as those associated with *in situ* testing in the exploratory shaft and with the implementation of currently proposed study plans. After all, it is an American proverb "if in doubt, don't."

APPENDIX C

APPENDIX C

**Material Regarding Whamonie Warm Spring Deposits, Prepared for
the NRC/NAS Panel and Presented by J. S. Szymanski During Field
Trip on April 23-25, 1991**

Issue - Is it reasonable to presume that, for the Nevada Test Site hydrosphere, the anticipated changes in position of the local water table may be measured in terms of hundreds of meters ?

- o Approximate altitude of the local land surface is ~ 4700 ft, Ekren and Sargent, 1965.
- o Approximate altitude of the regional water table is:
 - a) ~ 2420 ft, Domenico, 1972; and
 - b) closest wells encountered water table at ~ 2784 ft (Eleane Frm. - east) and ~ 3755 ft (upstream), Winograd and Thordarson, 1975.
- o Absence of any cover (including "pedogenic carbonates") suggests that age of the deposit may be as young as Holocene.
- o Presence of gypsum indicates that, the deposit parent fluids are:
 - a) deep-seated; and
 - b) were involved in geothermal circulations.

Basic data and main observations - the Whamonie CaSO_4 and CaCO_3 mounds.

The Paleospring Mound at Wahmonie. Although the only active warm springs close to Yucca Mountain are in Oasis Valley, there is a paleospring mound on the Nevada Test Site above Wahmonie, 28 km east of Yucca Mountain (fig. 1B). This spring mound consists of almost pure gypsum with minor calcite (fig. 14) and appears to represent deposits from an old hot-water discharge centered in the Wahmonie ore district, which was once mined for gold, silver, and copper (Bell and Larson 1982.). As at Tonopah, this hydrothermal system combines precious metals with sulfur-bearing mineralogy. Silica sinter, however, does not occur at Wahmonie. In its abundance of gypsum and lack of silica, the paleospring mound at Wahmonie is quite different from the fault-related calcite-silica deposits near Yucca Mountain.

From Vaniman et al., (1988).

Wahmonie Spring mounds - introductory information.

Figure C-3

At the Nevada Test Site, a deposit of gypsum, a few meters thick, was found near Whamonie. The outcrop of this deposit is located about 6 km northwest of Skull Mountain (Vaniman et al., 1988). The deposit occurs at the ground surface, in a topographic setting that precludes playa deposition, and in an area where the vadose zone is more than 500 m thick. It consists of almost pure gypsum with minor calcite, and leaves little doubt that it represents a warm-water paleo-spring discharge (Vaniman et al., 1988). Not much is known about the age of activity of this spring; field appearance of the outcrop, however, is one of a young deposit. The Whamonie paleo-spring mound occurs in a hydraulic setting where a) groundwaters are enriched in sulfate anion (Section 4.5.2); b) the intensity of flux of terrestrial heat is locally increased, (Section 4.3.2); c) the localized low-seismic velocity anomaly is present (Section 3.4.1); and d) hydrothermal convection system is locally developed (Sass et al., 1980). In this situation, it is reasonable to speculate that factors which are responsible for the Whamonie gypsum mound, may also be involved in the contemporary hydrology of Skull Mountain. In order to pursue such a hypothesis any further, however, it is necessary to establish the age of activity of the Whamonie mound. This age is a single most important data need that is required for a complete understanding of contemporary hydrology of the Skull Mountain area. A very modest field effort, consisting of mapping and trenching, is required to resolve this important issue.

From Szymanski, 1989.

Concurrent with drilling at Calico Hills, geophysical studies conducted at Wahmonie indicated that the granite, which occurs at the surface, would be only marginally large enough for a repository at the depth needed. These studies, plus surface mapping, also suggested that the granite within reasonable depths was probably altered by hydrothermal solutions (Smith et al., 1981; Hoover et al., 1982). In addition to the altered granite, local surface deposits from recent warm springs indicate upward seepage of groundwater, possibly from great depths. For these reasons, Wahmonie was eliminated from consideration in the spring of 1979.

From DOE, 1984 - based on USGS, 1980 letter.

Note:

- $\log (Q/K)$ data are from Kerrisk;
- $N = 48$ for hydraulic "source" regions, $N = 61$ for hydraulic "sink" regions; and
- Q - the ion activity product, K - the equilibrium constant for the solubility reaction.

Gypsum saturation index. Water samples from the Nevada Test Site hydraulic "sink" and "source" regions.

- o It is reasonable to conclude that, the parent fluids for the Whamonie gypsum mound have carried values of the gypsum saturation index $\log (Q/K) > 0$.
- o At the Nevada Test Site, none of the fluids sampled and analyzed have yielded positive values of the gypsum saturation index.

Conclusion: the parent fluids for the Wahmonie gypsum mound were upwelled from a depth measured in terms of kilometers.

Depth of fluid circulation, as indicated by the Whamonie Spring mounds.

- o An interpretation of the depth of fluid circulation is an important aspect of considerations regarding potential magnitudes of the tectonically induced rises of the water table.
- o Three independent lines of evidence indicate that, at the Nevada Test Site, the depth of groundwater circulation is large, namely:
 - i) presence of the fresh Whamonic mounds;
 - ii) the hydraulic "source" fluids residing in the Paleozoic carbonate complex carry abnormally high values of the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio, from 0.7118 to as much as 0.7149 . [a) in the pristine state, fluids equilibrated with the marine limestones of Early Paleozoic age should carry values of the $^{87}\text{Sr}/^{86}\text{Sr}$ ranging from 0.7082 to less than 0.7100, Faure (1986); b) the elevated values of the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio indicate that, at the Nevada Test Site, there is an upward fluid flux from the Pre-cambrian basement; and c) stratigraphic thickness of the Paleozoic carbonate complex exceeds ~ 2 km]; and
 - iii) some of the hydraulic "source" fluids residing in the Mesozoic non-carbonate rocks carry abnormally high values of the $\delta^{13}\text{C}$ ratio, e.g. a) sampling site #1 in Thirsty Canyon, $\delta^{13}\text{C} = -2.52$ per mil $_{\text{PDB}}$, White and Chuma (1978); b) the Yucca Mountain borehole USW H-3, $\delta^{13}\text{C} = -4.9$ per mil $_{\text{PDB}}$, Benson and McKinley (1985); and c) the Climax Stock perched waters, $\delta^{13}\text{C} = -3.54$ and -4.57 per mil $_{\text{PDB}}$, Isherwood et al. (1982). [a) the high values of the $\delta^{13}\text{C}$ ratio may be taken to indicate that fluids contain carbon derived through dissolution of the underlying Paleozoic carbonates; b) for sites with abnormally high values of the $\delta^{13}\text{C}$ ratio, the Paleozoic carbonates are known to occur at a depth ranging from ~ 1.5 to 4.0 km]; and
- o At the Nevada Test Site, the depth of groundwater circulation may reasonably be presumed to be in excess of ~ 5 km.

Interpretation of the depth of groundwater circulation for the Nevada Test Site.

Overall Conclusions

The Whamonie mounds provide direct and clear testimony that:

- i) at the Nevada Test Site, the tectonically induced (seismic and/or geothermal) rises of the local water table may be as large as a few hundred of meters;
- ii) at the Nevada Test Site, presence of the Benard-Rayleigh instabilities must be anticipated - confirms the results of dT/dz and $Ra - Rc$ analyses; and
- iii) downplaying potential magnitudes of the tectonic rises of the local water table [based on: a) beliefs, Dudley et al. (1989); b) reported global observations of hydrologic effects of earthquakes; and c) erroneous numerical simulations] is totally without merit.

APPENDIX D

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"500,000 - Year Stable Carbon Isotopic Record from Devils Hole, Nevada,"

by Tyler B. Coplen, Isaac J. Winograd, Jurete M. Landwehr, and Alan C. Riggs

"Stable Isotopic Evidence for a Pedogenic Origin of Carbonates in Trench 14 near Yucca Mountain, Nevada,"

by Jay Quade and Thure E. Cerling

**500,000-Year Stable Carbon Isotopic Record
from Devils Hole, Nevada**

Tyler B. Coplen,* Isaac J. Winograd, Jurate M. Landwehr, and Alan C. Riggs

500,000-Year Stable Carbon Isotopic Record from Devils Hole, Nevada

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Jurate M. Landwehr, Alan C. Riggs

The record of carbon-13 ($\delta^{13}\text{C}$) variations in DH-11 vein calcite core from Devils Hole, Nevada, shows four prominent minima near glacial terminations (glacial-interglacial transitions) V to II. The $\delta^{13}\text{C}$ time series is inversely correlated with the DH-11 oxygen isotope ratio time series and leads it by as much as 7000 years. The $\delta^{13}\text{C}$ variations likely record fluctuations in the $\delta^{13}\text{C}$ of dissolved inorganic carbon of water recharging the aquifer. How such variations are transported 80 kilometers to Devils Hole without obliteration by water-rock reaction remains an enigma. The record may reflect (i) global variations in the $\delta^{13}\text{C}$ of atmospheric CO_2 and, hence, the $\delta^{13}\text{C}$ of continental biomass or (ii) variations in extent and density of vegetation in the southern Great Basin. In the latter case, $\delta^{13}\text{C}$ minima at 414, 334, 246, and 133 thousand years ago mark times of maximum vegetation.

We have obtained a detailed and well-dated record of $\delta^{13}\text{C}$ variations in southern Great Basin ground waters for the period 60 to 566 thousand years ago (ka) from vein-calcite core DH-11 (1-3) from Devils Hole at the distal end of the Ash Meadows ground-water basin, Nevada (Fig. 1). This basin has an area greater than 12,000 km^2 and comprises a thick section (~ 100 to > 1000 m) of Paleozoic carbonate rocks that transmit water chiefly through fractures (4). The Spring Mountains, Pahrnagat Valley, and possibly the Sheep Range (Fig. 1) are the principal recharge areas (4, 5). Recharge also occurs by downward leakage from Tertiary volcanic and lacustrine aquifers (4). Devils Hole is near the center of

the principal discharge area, a 16-km-long, fault-controlled spring lineament at Ash Meadows (Fig. 1).

Vein-calcite samples from four locations in Devils Hole displayed $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ profiles versus time that are nearly identical (6). Our longest record, DH-11, is based on analysis of 285 samples (2) and shows periods of as much as tens of thousands of years during which the $\delta^{13}\text{C}$ of precipitating calcite was relatively constant (-1.6 to -1.8 per mil). These periods are separated by four prominent $\delta^{13}\text{C}$ troughs approximately centered on glacial terminations V through II, as delineated by the DH-11 $\delta^{18}\text{O}$ record (Fig. 2). In general (especially near terminations V to III), the $\delta^{13}\text{C}$ curve begins to decline at about the time the $\delta^{18}\text{O}$ curve reaches its minimum (glacial maximum). The $\delta^{13}\text{C}$ curve subsequently reaches its lowest (most negative) values at about the time that the $\delta^{18}\text{O}$ curve peaks (maximum interglacial conditions). Then, the $\delta^{13}\text{C}$

curve rapidly reverses direction (increasing $^{13}\text{C}/^{12}\text{C}$ ratio), whereas the $\delta^{18}\text{O}$ curve remains relatively high (at peak interglacial values) for 10,000 to 20,000 years before declining (7).

Although arising from distinct and independent geochemical processes, the DH-11 $\delta^{13}\text{C}$ profile (Fig. 2) is highly correlated ($r = -0.75$ with a $\delta^{18}\text{O}$ lag of 7000 years; Fig. 3) with the $\delta^{18}\text{O}$ time series (8). Spectral analyses (8) indicate robust peaks (in order of decreasing power) of 93,000, 40,000, 25,000, 23,000, and 17,000 years for the $\delta^{18}\text{O}$ time series and of 91,000, 40,000, 23,000, 28,000 and 18,000 years for the $\delta^{13}\text{C}$ data. Thus, obliquity and precession periodicities are evident in the DH-11 $\delta^{13}\text{C}$ record.

^{14}C and ^{13}C concentrations in dissolved inorganic carbon (DIC) from points along the flow path of the modern Ash Meadows ground-water basin indicate that extensive carbon exchange between water and aquifer rock occurs in the aquifer (9, 10). For example, ^{14}C and ^{13}C content vary from 79 pmc (percent modern carbon) and -9.5 per mil, respectively, in the Spring Mountains recharge waters, to 8.43 pmc and -7.6 per mil at Indian Springs (Fig. 1), 18 km down the hydraulic gradient. Ground water reaching Ash Meadows, tens of kilometers from Indian Springs, has average ^{14}C and $\delta^{13}\text{C}$ values of 2.1 pmc and -4.8 per mil, respectively (5). This carbon exchange is probably driven by episodes of calcite dissolution and precipitation as the ground-water flow alternates from depths of hundreds to more than a thousand meters below land surface and as temperature increases from 8° to 34°C (5). Carbon-isotope buffering tends to drive ^{13}C content of DIC toward equilibrium with aquifer carbonate rocks, which have $\delta^{13}\text{C}$ values ranging from -2 per mil (11) to perhaps several per mil more positive. This buffering also decreases the ^{14}C content of DIC, and ^{14}C ages are thousands of years too old (2).

With evidence for such extensive buffering, we expected to find a relatively featureless $\delta^{13}\text{C}$ profile in DH-11 calcite, not the impressive range and detail of the $\delta^{13}\text{C}$ fluctuations and the correlation with the $\delta^{18}\text{O}$ time series [which is assumed to be conservative in a low-temperature ($< 35^\circ\text{C}$) carbonate-rock aquifer]. In standard deviation units, many of the major $\delta^{13}\text{C}$ variations are larger than those of $\delta^{18}\text{O}$ (Fig. 3). That the DH-11 $\delta^{13}\text{C}$ record is responding in a significant manner to global climate is strongly suggested by these relations (Fig. 3).

We are unable to identify any process within the ground-water flow system that could generate the DH-11 $\delta^{13}\text{C}$ variations. We rule out several aquifer specific and general factors including (i) variation in water temperature or water level in Devils

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Hole (12), (ii) precipitation of calcite in isotopic disequilibrium (13), (iii) fluctuation in ground-water pH (14), (iv) occasional input of surface water to Devils Hole during storms (15), (v) variations in the fractions of flow derived from the principal recharge areas (16), (vi) varying ground-water residence times (17, 18), and (vii) varying growth rate of the vein calcite (19). We conclude that variations in the $\delta^{13}\text{C}$ profile must occur before carbon enters the aquifer.

How such variations are transmitted (albeit shifted to increasing $\delta^{13}\text{C}$ values) through the aquifer to Devils Hole >80 km distant without obliteration remains a major geochemical enigma. Nevertheless, we tentatively assume that the amount of buffering of each packet of water transversing the aquifer in the past was approximately constant in order for source variations to be transmitted.

Source variations of $\delta^{13}\text{C}$ may result from (i) variations in the relative abundance of plants in the principal recharge areas using C3 and C4 photosynthetic pathways; (ii) varying partial pressure of atmospheric CO_2 ($p\text{CO}_2$), which may affect the $\delta^{13}\text{C}$ of soil CO_2 and hence of DIC; (iii) varying $\delta^{13}\text{C}$ of atmospheric CO_2 and, hence, the $\delta^{13}\text{C}$ of continental biomass; and (iv) varying density and areal distribution of vegetation. Of these, the first does not seem to be important in the Ash Meadows system (20, 21). We also eliminate global $p\text{CO}_2$ variations because $\delta^{13}\text{C}$ values in DH-11 shift abruptly during peak interglacial times [133 to 118 ka (Fig. 2)] when atmospheric $p\text{CO}_2$ [derived from the Vostok ice core (22)] remained relatively constant.

Prominent $\delta^{13}\text{C}$ troughs centered approximately on terminations, as for DH-11 (Fig. 2), are also present in numerous $\delta^{13}\text{C}$ records of plankton in marine cores (23–26), including cores from regions with no upwelling where oceanic DIC might approach carbon isotopic equilibrium with atmospheric CO_2 (23, 24, 27) (Fig. 4). These marine $\delta^{13}\text{C}$ records (Fig. 4) also resemble the DH-11 curve in that the minima occur as much as several thousand years before $\delta^{18}\text{O}$ interglacial peaks [see (24)]. In both the marine and Devils Hole records, the descent from more positive $\delta^{13}\text{C}$ values begins as much as 20,000 years before the terminations, and the trough near termination III is not as distinct as those near terminations V, IV, and II. However, the amplitude of the $\delta^{13}\text{C}$ minimum centered on termination II is generally <1 per mil in the planktonic records (23, 24, 26) versus 1.3 per mil in DH-11. Moreover, because of reactions of water with Paleozoic carbonate detritus [$\delta^{13}\text{C}$ of –1 per mil (11)] in the soil, a 1.3 per mil change in the $\delta^{13}\text{C}$ of the DH-11 core would (for closed system conditions)

require a change in the $\delta^{13}\text{C}$ of atmospheric CO_2 of ~2.6 per mil. The largest shift in a marine record is 2.1 per mil (23), although bioturbation could have reduced the signal in some records (24).

Is there any evidence for an atmospheric

$\delta^{13}\text{C}$ teleconnection between the marine and DH-11 records? Krishnamurthy and Epstein (28) reported that $\delta^{13}\text{C}$ values in wood cellulose decreased by 4 per mil across termination I. They attributed this decline to a change in the $\delta^{13}\text{C}$ of atmospheric

Fig. 1. Index map of southcentral Great Basin, modified after (3). Major mountains are shaded heavy shading denotes altitudes of 2400 to 3600 m above sea level; light shading denotes altitudes of 1800 to 2400 m above sea level; ridge tops <1800 m above sea level are designated by name only. Heavy dashed and dotted line is the boundary of Ash Meadows ground-water basin (4); heavy dotted line is alternate eastern boundary (2). Arrows denote general direction of ground-water flow in aquifer as inferred from potentiometric maps (4). Dashed line is a major trough in potentiometric surface of carbonate aquifer. Solid line connecting Las Vegas and Lathrop Wells is U.S. Highway 95.

Fig. 2. Variations in $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ in core DH-11 during middle and late Pleistocene. Each dot represents an analysis; ages were assigned by interpolation between ^{230}Th - ^{234}U - ^{238}U dated sections of core (1, 2). Ages are minimum ages and need to be increased by ground-water travel times, which may be several thousands of years (2). Roman numerals denote approximate timing of glacial terminations (glacial-interglacial transitions) as in (2). Carbon isotope ratios are in per mil relative to VPDB [Vienna Pee Dee belemnite, defined by $\delta^{13}\text{C}_{\text{NBS19-VPDB}} = +1.95$ per mil (42)]. Oxygen isotope results are in per mil relative to VSMOW (Vienna Standard Mean Ocean Water) on a scale normalized such that SLAP (Standard Light Antarctic Precipitation) is –55.5 per mil (42). The $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ 1- σ analytic errors are 0.05 and 0.07 per mil, respectively. Dashed lines are based on analysis of five samples spanning the interval 57 to 24 ka of vein calcite sample DH-7 (6).

known to record accurately $\delta^{13}\text{C}$ variations of atmospheric CO_2 when not under salt stress. Their data display a prominent $\delta^{13}\text{C}$ minimum of ~ 1.1 per mil; however, the minimum is at ~ 18 ka, several thousand years before termination I. In contrast, analysis of the Byrd ice core (30) suggests that atmospheric $\delta^{13}\text{C}$ shifted only 0.3 per mil across termination I, and full glacial values were more negative than Holocene values. However, the ice core data between 20 and 10 ka are sparse.

The planktonic $\delta^{13}\text{C}$ records (23, 24, 26) show that the $\delta^{13}\text{C}$ trough centered on termination I is about half of that centered on termination II (Fig. 4). The DH-11 record does not extend to termination I, however, so a direct comparison cannot be made with the plant (28, 29) and ice core (30) records. The portion of the 1.3-per-mil shift (across termination II) in the DH-11 core that is attributable to global change may be clarified when CO_2 in 160,000- to 110,000-year-old ice from Greenland or Antarctica is analyzed for $\delta^{13}\text{C}$.

Fig. 3. Variation in $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ in core DH-11 in standard deviation units versus age. Curve for $\delta^{13}\text{C}$ is inverted. Age scale is same as in (2).

Fig. 4. Variation in $\delta^{13}\text{C}$ in planktonic foraminifera in three cores from non-upwelling, nutrient-poor portions of the western equatorial Pacific Ocean. Illustration from (24); vertical lines and Roman numerals have been added to show the location of SPECMAP (2) terminations V through I. Similar plots are published for cores from the west and east equatorial Atlantic (26) and from the southwestern Gulf of Mexico (23).

soil air varies inversely with vegetation density and extent (31). Declines in the DH-11 $\delta^{13}\text{C}$ profile (for example, from 355 to 335 ka in Fig. 2), which are coupled with ascending (warming) parts of the $\delta^{18}\text{O}$ curve, might therefore reflect increasing vegetation (and vice versa, as from 335 to 315 ka).

To address this hypothesis requires paleoecological—and related paleohydrologic and paleoclimatologic—data for the period 560 to 60 ka. Because such information is largely restricted to the past 30,000 years in the southern Great Basin, we assume that had the DH-11 $\delta^{13}\text{C}$ record extended to the Holocene, it would have shown (i) decreasing $\delta^{13}\text{C}$ values during the Wisconsin deglaciation, (ii) a $\delta^{13}\text{C}$ minimum as peak interglacial temperature was achieved (at ~ 11 ka), and (iii) sharply increasing $\delta^{13}\text{C}$ values between ~ 11 ka and the present.

Studies of packrat middens have demonstrated that vegetation ecotones in the southern Great Basin shifted as much as 1000 m downward during the last full glacial climate in comparison to modern plant occurrences (32). At such times, when mean annual temperatures are believed to have been at least 6°C cooler (33) and mean annual precipitation was 30 to 40% greater (34), pygmy-conifer woodlands extended (Fig. 1) to altitudes as low as 900 m (versus above 2000 m today); subalpine conifer forests typified by limber pine (*Pinus flexilis*) grew at altitudes as much as 600 m lower in the Sheep Range (33). It is thus likely that during full glacial times the highest several hundred to 1000 m of the Spring Mountains—presumably the major source of recharge—had alpine rather than the denser subalpine vegetation present today. Conversely, from the end of full glacial (21 ka) to peak interglacial (~ 11 ka) cli-

mate, which is not drought-tolerant *Pinus ponderosa*, co-existed with steppe shrubs until at least 13.2 ka. By 11.7 ka this community was replaced by a drought-tolerant pinon-juniper woodland. Concomitantly, pygmy-conifer woodland disappeared from all the intermediate altitude (<1800 m) ridges of the region (Fig. 1) by ~ 12 to ~ 10 ka, marking the end of a period of mesophytic vegetation.

Under the vegetative density hypothesis sparsest vegetation (in upland recharge areas) occurred during mid- to full glacial times (that is, ~ 390 to ~ 355 , ~ 315 to ~ 275 , ~ 200 to ~ 180 , and ~ 100 to ~ 20 ka; Fig. 2). Declining $\delta^{13}\text{C}$ values in the DH-11 core during deglaciations reflect the gradual revegetation of the highest (and wettest) 600 to 1000 m of the Spring Mountains, an area that, by analogy with the Sheep Range (33, 34), likely had only alpine or sparse subalpine vegetation during full glacial climates. Densest vegetation was attained at the beginning of the last four interglacials at 414, 334, 246, and 133 ka. The sharp reversals and increase in $\delta^{13}\text{C}$ that began just after these times (Fig. 2) are taken to reflect a sudden and dramatic reduction in vegetative density or extent in response to high interglacial temperature and possibly changes in seasonality of precipitation (35). This hypothesized reduction in vegetation at the beginning of interglacials (several thousand years before a decrease in $\delta^{18}\text{O}$) may account for why the $\delta^{13}\text{C}$ time series leads the $\delta^{18}\text{O}$ time series. The $\delta^{13}\text{C}$ reversal may also reflect deforestation of the intermediate altitude ridges of the region [those less than 1800 m, for example, Desert, Pinwater, Spotted, and Specter ranges (Fig. 1)], provided such ridges contributed significantly to full glacial recharge. This deforestation was completed between ~ 12 and ~ 10 ka (33), and presumably at equivalent times in earlier glacial cycles. According to this hypothesis, the sharp $\delta^{13}\text{C}$ minimum at 133 ± 2 ka would indicate that the retreat of the North American ice sheets was essentially complete 5000 years before the exceptional high in Northern Hemisphere summer insolation at 128 ± 1 ka (36).

Decreasing $\delta^{13}\text{C}$ values during deglaciations may also be explained by glacial meltwater discharges into the Gulf of Mexico, which might have increased summer precipitation over the southwestern United States by strengthening the Bermuda high and shifting it westward (37). In addition, the sharp reversals in $\delta^{13}\text{C}$ (particularly at 334, 246, and 133 ka) might reflect (38,

**Stable Isotopic Evidence for a Pedogenic Origin of
Carbonates in Trench 14 near Yucca Mountain, Nevada**

JAY QUADE AND THURE E. CERLING

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Layered carbonate and silica encrust fault fractures exposed in Trench 14 near Yucca Mountain, site of the proposed high-level nuclear waste repository in southern Nevada. Comparison of the stable carbon and oxygen isotopic compositions of the fracture carbonates with those of modern soil carbonates in the area shows that the fracture carbonates are pedogenic in origin and that they likely formed in the presence of vegetation and rainfall typical of a glacial climate. Their isotopic composition differs markedly from that of carbonate associated with nearby springs. The regional water table therefore remained below the level of Trench 14 during the time that the carbonates and silica precipitated, a period probably covering parts of at least the last 300,000 years.

ONE OF THE CHIEF ASSUMPTIONS used to justify locating a high-level nuclear waste repository at Yucca Mountain is that the buried waste will remain well within the unsaturated zone for the next several hundreds of thousands of years, regardless of even large changes in water-table elevation in response to climate change. A serious challenge to the validity of this assumption was the discovery of a complex network of carbonate and silica fillings in the fractures associated with several faults bordering Yucca Mountain. The best known and the most controversial of these fracture fillings are located in Trench 14, which crosses the Bow Ridge fault (Fig. 1). The thick, well-layered, and well-indurated nature of the fillings caused immediate concern because of similarity in these respects to vein cements and travertines associated with springs (1) and because of the apparent Quaternary age of some of the carbonate (2). A spring origin would imply that there was a rise in the regional water table, presumably during glacial maxima, to at least the level of Trench 14. Trench 14 is about 150 m above the level of the proposed repository and 400 m above the water table. A return to glacial hydrologic conditions might then result in ground-water flooding of the repository and rapid transport of radionuclides to nearby discharge points, such as Trench 14.

There has been a substantial effort at

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establishing the origin of the fracture cements in Trench 14 (1). An alternate hypothesis to a spring origin is that the carbonates and silica formed in soils in the vadose zone for hundreds of thousands of years (2). In this case, meteoric water infiltrating through the fractures deposited carbonate and silica as a normal part of desert soil formation.

To test these two hypotheses, we compared the carbon and oxygen isotopic data from Trench 14 to those in modern desert soils. We sampled soils in settings where the water table has remained tens to hundreds of meters below the surface during pedogenesis and therefore where ground water has played no role in carbonate formation. We took special care to sample soils younger than 7000 years old in order to establish the relation between the isotopic composition in soil carbonate and modern vegetation and rainfall (3). Pack-rat midden and pollen evidence shows that the distribution of vegetation has not greatly changed in the region during that span (4). We sampled soils displaying weak Stage I morphology (5), a degree of development consistent with a mid-to-late Holocene age. Five accelerator dates on thin carbonate coatings from three representative soils yielded ages between 3820 and 820 years (Table 1) (6).

The $\delta^{13}\text{C}$ content of modern soil carbonate varies substantially with elevation (Fig. 2) because of variations in (i) the isotopic composition of desert plants and (ii) the proportion of atmospheric CO_2 in the des-

ert soil atmosphere (3). Plants fractionate C along three differing metabolic pathways: C_3 , C_4 , and CAM. The C_3 plants, including conifers, mountain shrubs, and some grasses, make up virtually all of the biomass at higher elevations in the southern Great Basin. The $\delta^{13}C$ of these plants average -24 to -25 per mil at high-elevation sites, and coexisting soil carbonate has a $\delta^{13}C$ higher than in this biomass by 14 to 16 per mil (7). At lower elevations, C_4 and CAM (8) plants, as well as many C_3 shrubs and herbs, are present; here, C_4 plants are the most abundant, and they have an average $\delta^{13}C$ of -13 per mil. A larger proportion of atmospheric CO_2 , with an average $\delta^{13}C$ of -6 per mil, is present deep in soils at low elevations because of lower plant respiration rates. This C reservoir further increases the $\delta^{13}C$ of soil CO_2 —and therefore also of soil carbonate—that is derived from the mix of C_3 and C_4 plants. Holocene-age soil carbonates from each of the major vegetation zones therefore display a distinct $\delta^{13}C$ signature (Fig. 2). For example, the $\delta^{13}C$ of soil carbonates average -9 per mil in the ponderosa pine zone (>2400 m) and -7.4 ± 0.8 per mil (15 samples) in the pinyon-juniper-sage zone (1800 and 2300 m). Near sea level in Death Valley, the $\delta^{13}C$ of soil carbonates is -2 to -2 per mil, reflecting the contribution of C_4 creosote and C_4 desert holly shrubs, as well as some mixing of atmospheric CO_2 deep in the soil (3). The $\delta^{18}O$

Fig. 1. The fracture system filled with layered calcite and silica, exposed on the south wall, east end of Trench 14. The trench wall is about 3.5 m high.

Table 1. The ^{14}C dates (in ^{14}C years before present) by accelerator mass spectrometry on pedogenic carbonate from three (SM-2, 3, 4) soils in the Spring Mountains, 100 km south of Yucca Mountain (6). For each sample, we scraped 50 to 100 mg of material from clast undersides, taking care not to include any carbonate from the host clast or from older cements. Further details on soil sites, designated by these field numbers, is in (3).

Lab no.	Field no.	Age	Soil depth (cm)
AA-3697	SM-3(B)	1210 $\pm$ 50	15
AA-3698	SM-2A	1900 $\pm$ 60	Surface
AA-3699	SM-2B	3820 $\pm$ 55	10 to 15
AA-3700	SM-2C	1785 $\pm$ 65	30 to 40
AA-3701	SM-4	820 $\pm$ 50	50 to 60

composition of soil carbonate also decreases with elevation (Fig. 3), irrespective of parent material, primarily because the $\delta^{18}O$ composition of rainfall also decreases with increasing elevation. Evaporation or differing penetration of meteoric water seasonally also may influence the $\delta^{18}O$ of pedogenic carbonates (3).

We sampled five Holocene-age soils in the vicinity of Trench 14. The local vegetation is dominated by C_3 shrubs such as black bush (*Coleogyne ramosissima*) and Nevada joint fir (*Ephedra nevadensis*); creosote (*Larrea divaricata*), burrobrush (*Ambrosia dumosa*), and twin fruit (*Menodora spinescens*) are less abundant. Shadscale (*Atriplex confertifolia*), a C_4 shrub, is also common. All the herbs, and some of the grasses (mainly *Stipa* sp. and *Oryzopsis hymenoides*), are C_3 plants. Fluffgrass (*Leroneuron pulchellum*), a C_4 grass, grows in the early summer. This mix of C_3 and C_4 plants, and moderate plant respiration rates, produces $\delta^{13}C$ values in soil carbonate of -4.2 to -7.1 per mil (14 samples) below a depth of 50 cm. The $\delta^{18}O$ (PDB) of the same carbonates nearly all fall between -7 and -10.6 per mil.

We analyzed 22 samples from most of the major veins exposed in Trench 14. The $\delta^{13}C$ of all but one of samples ranges from -6.3 to -7.7 per mil (average -7.0 ± 0.4 per mil, Fig. 2). The Trench 14 carbonates have low $\delta^{13}C$ values compared to those in nearby Holocene-age soil carbonate: these low values indicate that the Trench 14 carbonates did not form in equilibrium with the modern vegetation in the area. However, the C isotope data do overlap with values from modern soils about 750 m above the site, where cover is dominated by pinyon (*Pinus monophylla*), juniper (*Juniperus osteosperma*), and sagebrush (*Artemisia tridentata*). Samples from this vegetation zone yielded a $\delta^{13}C$ average of -7.4 ± 0.8 per mil. A single sample from Trench 14 yielded a result of -4.6 per mil, a value that is

Fig. 2. The $\delta^{13}C$ (PDB) of pedogenic carbonate in modern soils developed on volcanic parent materials along elevation transects in the southern Great Basin. The linear regression (shown) for 37 soil analyses is: elevation (meters) = $554.4 + 103.8(\delta^{13}C - 17.498) \pm 18.3$, where $\delta^{13}C$ is the $\delta^{13}C$ (PDB) of soil carbonate. The $\delta^{13}C$ compositions of all but one sample from Trench 14 are lower than that for carbonate in Holocene-age soils in the vicinity but overlap with the $\delta^{13}C$ range of soil carbonate ~750 m higher on the transects.

observed in modern soils nearby. The other 21 samples from Trench 14 display a narrow range of $\delta^{18}O$ (PDB), between -10.5 and -11.8 per mil (average of -11.3 ± 0.3 , Fig. 3). Again, this is close to the values for modern soil carbonate in the pinyon-juniper-sage zone (average -11.5 ± 1.4 per mil) and is unlike nearby modern soils at -9.8 ± 1.2 per mil (9).

Evidence from fossil pack-rat middens indicates that vegetation zones were displaced about 1,000 m downward during the last full-glacial period (16,000 to 19,000 years ago) (4). Juniper, sagebrush, and pinyon then dominated the vegetation at Yucca Mountain. Our evidence indicates that the Trench 14 carbonates, if formed in soils, would have precipitated in equilibrium with this plant assemblage during the glacial maxima. The $\delta^{18}O$ of ground water in the region during glacial times was 1.5 to 2.0 per mil less than that of modern ground water (10, 11); this is close to the 1.5-per-mil average difference we observe between carbonates in Trench 14 and nearby modern soils.

A further test of a soil versus spring origin comes in comparing the C isotopic composition of coexisting carbonate and occluded organic matter. The difference should be 14 to 16 per mil in soils with relatively high respiration rates, including those found under pinyon-juniper-sage cover (7). The Trench 14 carbonates contain 0.12 to 0.18% organic C (Table 2). The average difference between carbonates and occluded organic matter was found to be 14.6 ± 0.2 per mil (3 samples), which supports a pe-

Table 2. The $\delta^{13}\text{C}$ (PDB) of coexisting carbonate and occluded organic (org.) matter from Trench 14. Roughly 30-g samples of carbonate were pyrolyzed at 650°C for 1 hour in covered crucibles, thus converting organic matter to pure carbon. Carbon was then concentrated by hydrolysis of the carbonate with 3N HCl.

Sample	Org. C (per cent)	$\delta^{13}\text{C}$ carbonate (per mil)	$\delta^{13}\text{C}$ org. matter (per mil)	Difference (per mil)
YM-14-8B	0.18	-6.3	-21.2	14.9
YM-14-11B	0.13	-7.0	-21.6	14.6
YM-14-12B	0.12	-7.3	-21.7	14.4

*By weight.

ogenic origin for the vein fillings in Trench 14. The spring carbonates we collected from veins at Ash Meadows (see below) did not contain sufficient occluded organic matter for analysis.

Uranium-trend dates and other geologic evidence (2, 12) suggest that carbonate in several fractures exposed in Trench 14 accumulated during the last 300,000 years, but that other fractures were filled earlier. As glacial periods have occurred about every 100,000 years (10), cementation events in the fractures could have occurred in one or more of many glacial periods during the latter half of the Pleistocene, and possibly earlier.

How do the isotope results on carbonates from Trench 14 compare to those expected for carbonates of unequivocal spring origin? To answer this question we considered two situations that account for the different types

Fig. 3. The $\delta^{18}\text{O}$ (PDB) of pedogenic carbonate in modern soils developed on volcanic parent materials along elevation transects in the southern Great Basin. The linear regression (shown) for 39 soil analyses is: elevation (in meters) = $(-149.0 \pm 207.7) - (173.7 \pm 18.5)z$, where z is the $\delta^{18}\text{O}$ (PDB) of soil carbonate. On average, Trench 14 carbonates have $\delta^{18}\text{O}$ values consistent with soil carbonate now ~790 m above the site in the lower pinyon-juniper-sagebrush zone, the same setting indicated by the C isotopes in Trench 14.

of spring conditions found in the region: (i) spring discharge from the regional water table in fractured, transmissive bedrock; and (ii) cool spring discharge fed by local perched water.

Vein carbonate associated with springs that discharge from the regional water table have been analyzed in the Ash Meadows area south of Yucca Mountain (10, 13). Carbonates there and at Trench 14 fill major extensional fractures within a few meters of the surface, from which, in the case of Ash Meadows, ground water discharges directly from a major aquifer. Although different ground-water systems underlie the two areas, the age, temperature, and most aspects of the isotope chemistry of the two systems are alike. The $\delta^{18}\text{O}$ (SMOW) of ground water at Ash Meadows is nearly identical to that of the ground water underlying Trench 14, at -13 to -13.5 per mil (10, 11, 14). However, the $\delta^{13}\text{C}$ of dissolved inorganic carbon in Ash Meadows ground water is greater by 3 to 4 per mil in comparison to ground water under Trench 14, probably because of exchange with Paleozoic carbonate rocks along the flow path.

The $\delta^{18}\text{O}$ of spring carbonates from Devil's Hole, located near to Ash Meadows, is about -14.5 and -16.5 per mil (PDB) for interglacial and glacial climates, respectively (10) (Fig. 4). These values are appreciably lower than values for Trench 14 carbonates because of the low $\delta^{18}\text{O}$ ratios of the ground waters and high temperatures of the springs. The $\delta^{18}\text{O}$ values of carbonates in Trench 14 require temperatures of formation of $\leq 15^\circ\text{C}$ (15), if the inferred $\delta^{18}\text{O}$ range of ground water during the last 250,000 years (10, 11) is correct. Spring discharge from the regional aquifer at such low temperatures is highly unlikely at Trench 14 in view of the geologic setting. Temperatures in springs at Ash Meadows-Devil's Hole and in ground water under Trench 14 (14) generally exceed 28°C because of deep circulation of ground water, and these temperatures are unaffected by short-term ($< 300,000$ year) climate fluctuations (10). Temperatures as low as 20°C are known from minor springs fed by deeply circulated water at Ash Meadows. But in these springs, water passes slowly to the surface through aquitards composed of fine-grained Pliocene basin-fill, probably after leakage from the more transmissive carbonate aquifer below. The slow circulation apparently causes spring water to partially or wholly equilibrate with near-surface temperatures (16). No fine-grained strata have been observed in the Trench 14 area. The transmissive nature of the fractured volcanic rocks and fault system at Trench 14 would have led to rapid ascent of ground water, as in the examples from Ash Meadows, where

Fig. 4. The $\delta^{18}\text{O}$ (PDB) versus $\delta^{13}\text{C}$ (PDB) of carbonates from Holocene soils (□), Trench 14, and springs. In all, 21 of the 22 analyses from Trench 14 fall within the isotopic field for Holocene-age soil carbonate found in the pinyon-juniper-sagebrush zone on volcanic parent material. A single result lies in the field defined by modern soils in the vicinity. Spring carbonates from Ash Meadows (13) and Devil's Hole (10) plot completely outside the observed range for carbonates in modern soils and in Trench 14.

springs flowing from bedrock have temperatures $\geq 28^\circ\text{C}$ irrespective of discharge rate (16).

The $\delta^{13}\text{C}$ of spring carbonate at Ash Meadows and Devil's Hole ranges from -1.5 to -2.9 per mil (17) (Fig. 4). This carbonate precipitated in isotopic equilibrium with HCO_3^- in ground water that had a $\delta^{13}\text{C}$ of -4 to -5 per mil. The HCO_3^- in ground water under Yucca Mountain is much more variable, ranging between -11 and -2 per mil (14). If such a broad range of spring water compositions did produce over a long period the narrow range of $\delta^{13}\text{C}$ values observed in Trench 14 carbonates, it would be highly fortuitous.

The overriding difficulty with the second situation described above, spring discharge of local perched water, is that the hydrogeologic setting is generally wrong. Springs in perched water systems normally display a large upgradient catchment area and an aquitard unit that crops out at the point of discharge. Perched water in the region is typically found above and discharges from strata with a low permeability, such as zeolitized air-fall tuffs (16). In contrast, the rocks at Trench 14 are highly fractured ash-flow tuff and permeable alluvium; the nearest thin ash-fall aquitards are several hundred meters below Trench 14. However, the possibility of a perched water setting cannot be refuted by our isotopic results. The water in perched zones could have originated locally as precipitation and therefore had a similar oxygen isotopic composition to soil water. The HCO_3^- in perched

water may also be in equilibrium with locally derived plant CO_2 , as HCO_3^- is in soil water.

Other evidence also supports a pedogenic origin for the Trench 14 carbonates. The morphology (2) and petrography of the carbonates and silica fillings are consistent with a soil origin, as is micromorphological, clay mineralogical, trace element (18), and isotope tracer (19) evidence. Oxygen isotopes from the silica cements indicate that the temperatures of formation were $\sim 15^\circ\text{C}$, consistent with that in a pedogenic environment (2).

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8. CAM plants, which include cactus and most yucca, are present at lower elevations but don't exceed 10% of the biomass at any site. They average -18 per mil in $\delta^{13}\text{C}$.
9. We observed that the $\delta^{13}\text{C}$ (PDB) and the $\delta^{18}\text{O}$ (PDB) of carbonates appear to be constant below about 40-cm depth in modern soils at this general elevation (3). Above 40 cm, mixing with the atmospheric CO_2 causes progressive enrichment up to the soil surface. All the Trench 14 samples come from 0.9 to 3.4 m below the surface, and display no systematic trends with depth. All our modern soil carbonates that we used to fingerprint isotopically each of the vegetation zones come from 0.5 to 2.2 m depth.
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20. We thank M. Mifflin, I. Winograd, and L. Benson for their useful discussions and help, and R. F. Johnson for his logistical assistance. Analyses were performed at the University of Utah and Yale University. This work was supported primarily by the Nevada Nuclear Waste Project Office, and by National Science Foundation grant BNS-8703304.

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APPENDIX E

APPENDIX E

Isotopic Data Used to Perform Interpretations of the Origin of the Trench 14 Veins and the Veins from Subsurface

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Sample No.	Elevation (m)	$\delta^{13}\text{C}_{\text{POB}}$	$\delta^{18}\text{O}_{\text{POB}}$
SM-1a	840	-2.0	-9.0
SM-1b	840	-1.9	-7.8
SM-1c	840	-1.2	-7.8
SM-2b	1550	-3.1	-8.8
SM-2b	1550	-4.0	-11.2
SM-2c	1550	-5.2	-12.3
SM-3a	1950	-5.4	-11.2
SM-3b	1950	-5.0	-11.6
SM-3c	1950	-5.8	-10.4
SM-3(B)a	1900	-4.3	-11.0
SM-3(B)b	1900	-3.6	-9.4
SM-3(B)c	1900	-4.8	-9.2
SM-4a	2270	-8.1	-13.1
SM-4b	2270	-6.0	-8.2
SM-4c	2270	-5.8	-9.2
SM-4d	2270	-7.3	-11.4
SM-5a	2490	-8.62	-12.2
SM-5b	2490	-8.4	-12.2
SM-5b-rerun	2490	-8.4	-12.5
SM-5c	2490	-8.8	-11.5
SM-6a	2740	-9.11	-12.2
SM-6b-1	2740	-8.08	-11.8
SM-6b-2	2740	-8.7	-12.6
SM-6c	2740	-8.0	-11.7
TC-1a	300	3.4	-2.8
TC-1b	300	4.1	-2.6
TC-1c	300	2.5	-5.6
GM-1a	1830	-8.3	-11.0
GM-1a-1m	1830	-6.9	-11.6
GM-1c	1830	-6.5	-10.4
GM-1d	1830	-7.2	-11.0
GM-2a-90cm	1575	-6.3	-9.4
GM-2b	1575	-5.8	-8.8
GM-2c	1575	-5.5	-9.7
GM-3a	1160	-2.0	-6.7
GM-3b	1160	-0.9	-7.1
GM-3c	1160	-1.8	-7.0
GM-4a	900	-2.4	-6.8
GM-4b	900	-1.3	-7.5
PaM-1a	300	-1.7	-0.4
PaM-1b	300	-0.4	-3.5
PaM-1c	300	0.9	-6.2
PiVM-1a	1675	-6.1	-11.1
PiVM-1d	1675	-6.1	-11.9
PiVM-2a	2130	-8.1	-11.2
PiVM-2b	2130	-7.5	-10.0
PiVM-2c	2130	-7.7	-11.2

Figure E-1. Carbon and oxygen isotopic data. Cobble encrustations - region around the Nevada Test Site. From Quade et al., 1988.

Sample ID	Location	$\delta^{18}\text{O}_{\text{SMOW}}$	$\delta^{13}\text{C}_{\text{PDB}}$	Source
32 d	Whamonie	19.85	-4.4	Harmon (1993)
24 d	Crater Flat	19.90	-2.8	Hill and Schluter (1994)
HD-164-2	"	18.90	2.8	"
HD-164-2A	"	18.80	-2.8	"
HD-166-1	"	20.5	-1.3	"
HD-166-1	"	20.5	-1.3	"
HD-167A	"	20.1	-2.1	"
HD-168-1A	"	21.50	-1.5	"
HD-169-1	"	21.30	-1.1	"
HD-169-1A	"	21.40	-1.0	"
HD-443-4A	Whamonie	18.10	-5.0	DOE. (1993)
HD-443-5A	"	23.60	-4.8	DOE. (1993)

Figure E-2. Isotopic character of carbon and oxygen incorporated in known spring deposits from around Yucca Mountain.

Figure E-3. Location map - boreholes discussed in the text.

HD #	Drill Hole	SPC #	Depth (ft)	Tcrush (C)	Thomog (C)
HD-318	G1	00016017	425.5	95,90,-90	
HD-319	G1	00016018	491.8	-60,60	
HD-322	G1	00016021	669.2	-100,99.5,76,92	81
HD-325	G1	00016024	959.8	94,-82	
HD-326	C1	00016025	1029.0	82,82,82,-82	
HD-338	G	00016037	3588.5		99,101,114
HD-343	G1	00016042	3867.3		74,86
HD-348	G1	00016047	5348.6		87,96,92,91
HD-359	G2	00016060	858.2		57,59
HD-368	G2	00016068	1137.6		81>72
HD-369	G2	00016069	1169.8		104,103
HD-274	G2	00004688	1562.2	94,92	
HD-579	G2	00016083	4162.8		78,-80,82,80,79
HD-582	G2	00016086	4912.8		org(?) -240-260
HD-583	G2	00016087	5107.3		215,216,242-245

Note: The corresponding stable isotope data are provided in Figures E-11 through E-14.

Figure E-4. Homogenization temperatures of fluid inclusions in calcitic veins. Yucca Mountain. From D.O.E., 1993.

Depth [m]	Homogenization temp. of fluid inclusions [°C]	Source
Borehole USW G-2		
1640	from 94 to 115	Bish, 1989
1756	147	do
1773	from 202 to 239	do
Borehole USW G-3		
1463	97	do
Borehole USW GU-3		
32	from 101 to 227	do
130	from 125 to 170	do

Note:

"Other authigenic minerals in the Topopah Spring unit of USW GU-3 include silica, calcite, and fluorite. Calcite and fluorite are very rare, although sparry calcite is concentrated in veins and as a cement in fault breccia at and above 525 ft depth, and it occurs with fluorite in a fracture at 1027 ft. depth. Sparry calcite in fault breccia at 482 ft. depth is coated by a later growth of fine-grained anhedral calcite, which suggests temporary saturation within the breccia during the sparry-growth cementation, followed by vadose calcite growth" Vaniman et al., 1984.

Figure E-5. Homogenization temperatures of fluid inclusions in calcitic veins. Yucca Mountain. From Szymanski, 1992.

Locality	HG#	Depth, m	$\delta^{18}\text{O}$ in Opal, ‰ SMOW	$\delta^{18}\text{O}$ in Calcite, ‰ SMOW	Δ , ‰	Source
Pull Apart fault	-	0	28,17	23,36	4,81	Hammon, 1993
Bare Mountain	-	0	28,19	20,27	7,92	- " -
Wailing Wall	-	0	28,17	19,84	8,33	- " -
Busted Butte	-	0	29,91	20,23	9,68	- " -
USW G-2	351B	92,2	26,6	20,6	6,00	YMP, 1993
USW G-2	355A	236,7	20,6	17,9	2,70	- " -
USW G-2	356A	240,7	21,7	19,0	2,70	- " -
USW G-2	358A	257,8	24,1	17,4	6,70	- " -
USW G-2	362A	280,2	21,2	18,9	2,30	- " -
UE-25 A-5	926A	85,2	21,9	17,8	4,10	- " -
UE-25 A-5	929A	92,2	23,3	17,9	5,40	- " -

Note:

a) For a 34 -93 °C range, $\Delta_{s-c} = 0.74 (10^6 / T^2) - 0.69$, and

b) For a 200 - 500 °C range, $\Delta_{s-c} = 0.60 (10^6 / T^2)$.

Figure E-6. Δ_{c-o} values, calcite-opal pairs from Yucca Mountain. From TRAC -N.A., 1994.

Figure E-7. Well USW G-1. Temperature profile. From Sass et al., 1987.

Figure E-8. Well USW G-2. Temperature profile. From Sass et al., 1987.

Figure E-9. Well USW G-3. Temperature profile. From Sass et al., 1987.

Figure E-10. Well UE-25a No. 5. Temperature profile. From Sass et al., 1987.

CALCITE/SILICA - CARBONATE STABLE ISOTOPES

Accept?	HD #	Locality	Depth : m	Estn Log #	d13C pdb	d18O smow	SPC #
☒ Yes	HD-322C	USW G-1	204.0	46C-2	-4.6	14.8	00016021
☒ Yes	HD-322C	USW G-1	204.0	48C-5	-5.2	15.0	00016021
☒ Yes	HD-325A	USW G-1	292.5	46C-3	-7.3	15.8	00016024
☒ Yes	HD-325A	USW G-1	292.5	48C-6	-7.3	15.8	00016024
☒ Yes	HD-326A	USW G-1	313.6	39C-6	-6.8	15.6	00016025
☒ Yes	HD-326A	USW G-1	313.6	48C-8	-6.8	15.6	00016025
☒ Yes	HD-327A	USW G-1	310.1	46C-5	-8.2	16.6	00016026
☒ Yes	HD-327A	USW G-1	310.1	48C-10	-7.9	16.6	00016026
☒ Yes	HD-327B	USW G-1	310.1	39C-10	-4.2	15.0	00016026
☒ Yes	HD-327B	USW G-1	310.1	49C-1	-4.5	14.8	00016026
☒ Yes	HD-327C	USW G-1	310.1	49C-2	-7.8	16.2	00016026
☒ Yes	HD-328A	USW G-1	345.7	49C-3	-7.5	16.3	00016027
☒ Yes	HD-328A	USW G-1	345.7	80C-4	-7.6	16.3	00016027
☒ Yes	HD-328B	USW G-1	345.7	49C-4	-7.1	15.5	00016027
☒ Yes	HD-329A	USW G-1	353.7	40C-2	-1.7	14.6	00016028
☒ Yes	HD-329A	USW G-1	353.7	49C-6	-1.6	14.2	00016028
☒ Yes	HD-330A	USW G-1	365.4	40C-3	-7.6	16.5	00016029
☒ Yes	HD-330A	USW G-1	365.4	49C-7	-7.5	16.5	00016029
☒ Yes	HD-330B	USW G-1	365.4	49C-8	-6.5	16.4	00016029
☒ Yes	HD-330B	USW G-1	365.4	52C-8	-6.4	16.8	00016029
☒ Yes	HD-331A	USW G-1	385.7	49C-9	-5.4	12.7	00016030
☒ Yes	HD-331A	US	385.7	80C-6	-5.5	13.1	00016030
☒ Yes	HD-334A	US	895.3	50C-1	4.4	11.8	00016033
☒ Yes	HD-334A	USW G-1	895.3	80C-7	4.2	11.9	00016033
☒ Yes	HD-336A	USW G-1	1090.3	50C-2	1.9	10.3	00016035
☒ Yes	HD-337A	USW G-1	1090.7	50C-3	1.7	10.0	00016036
☒ Yes	HD-337B	USW G-1	1090.7	50C-4	1.9	10.3	00016036
☒ Yes	HD-338A	USW G-1	1093.8	50C-5	2.1	10.0	00016037
☒ Yes	HD-339A	USW G-1	1109.8	50C-6	3.3	11.9	00016038
☒ Yes	HD-339A	USW G-1	1109.8	52C-9	3.3	11.9	00016038
☒ Yes	HD-340A	USW G-1	1112.7	50C-7	2.1	10.2	00016039
☒ Yes	HD-341A	USW G-1	1120.7	50C-9	2.3	10.1	00016040
☒ Yes	HD-341A	USW G-1	1120.7	51C-5	2.3	10.0	00016040
☒ Yes	HD-342D	USW G-1	1164.9	50C-8	3.3	11.3	00016041
☒ Yes	HD-343B	USW G-1	1178.8	50C-10	2.3	10.7	00016042
☒ Yes	HD-343C	USW G-1	1178.9	51C-6	1.9	11.0	00016042
☒ Yes	HD-343D	USW G-1	1178.8	51C-7	2.5	10.2	00016042
☒ Yes	HD-344B	USW G-1	1187.4	51C-8	1.4	9.6	00016043
☒ Yes	HD-345B	USW G-1	1427.0	51C-9	0.5	3.9	00016044
☒ Yes	HD-345B	USW G-1	1427.0	80C-8	0.4	4.0	00016044
☒ Yes	HD-346A	USW G-1	1550.0	51C-10	-0.2	3.4	00016045
☒ Yes	HD-346A	USW G-1	1550.0	80C-9	-0.3	3.4	00016045
☒ Yes	HD-347A	USW G-1	1618.8	52C-2	0.1	9.5	00016046
☒ Yes	HD-347A	USW G-1	1618.8	80C-10	0.0	9.6	00016046
☒ Yes	HD-348A	USW G-1	1630.3	52C-3	0.1	2.4	00016047
☒ Yes	HD-348B	USW G-1	1630.3	52C-4	0.4	4.2	00016047
☒ Yes	HD-349A	USW G-1	1636.3	52C-5	0.4	1.3	00016048
☒ Yes	HD-350A	USW G-1	1654.6	52C-5	-0.4	6.9	00016050
☒ Yes	HD-350A	USW G-1	1654.6	52C-6	-0.4	6.9	00016050
☒ Yes	HD-350B	USW G-1	1644.6	53C-6	0.2	8.2	00016050

**Figure E-11. Borehole USW G-1, stable isotope data, calcitic veins.
From D.O.E., 1993.**

CALCITE/SILICA - CARBONATE STABLE ISOTOPES

Accept?	HD #	Locality	Depth : m	Extm Log #	d13C pdb	d18O smow	SFC #
☑ Yes	HD-280A	USW G-1	885.4	347-C-6	-9.5	14.2	00004694
☑ Yes	HD-280A	USW G-1	885.5	34C-9	-9.4	14.4	00004694
☑ Yes	HD-280A	USW G-1	885.4	46C-8	-9.5	14.0	00004694
☑ Yes	HD-317A	USW G-1	97.7	34C-10	-5.5	17.4	00016016
☑ Yes	HD-317A	USW G-1	97.7	46C-9	-5.7	16.9	00016016
☑ Yes	HD-317B	USW G-1	97.7	35C-1	-6.1	17.1	00016016
☑ Yes	HD-317B	USW G-1	97.7	46C-10	-6.1	17.1	00016016
☑ Yes	HD-318A	USW G-1	129.7	35C-2	-4.8	14.6	00016017
☑ Yes	HD-318A	USW G-1	129.7	47C-1	-4.9	14.5	00016017
☑ Yes	HD-318A	USW G-1	129.7	47C-1	-4.9	14.4	00016017
☑ Yes	HD-318B	USW G-1	129.7	38C-2	-6.8	16.2	00016017
☑ Yes	HD-318B	USW G-1	129.7	47C-2	-6.8	16.0	00016017
☑ Yes	HD-319A	USW G-1	149.9	38C-3	-5.4	14.6	00016018
☑ Yes	HD-319A	USW G-1	149.9	38C-3	-5.4	14.7	00016018
☑ Yes	HD-319A	USW G-1	149.9	38C-3	-5.4	14.7	00016018
☑ Yes	HD-319A	USW G-1	149.9	47C-4	-5.4	14.3	00016018
☑ Yes	HD-319B	USW G-1	149.9	38C-4	-2.4	14.2	00016018
☑ Yes	HD-319B	USW G-1	149.9	80C-3	-2.4	14.3	00016018
☑ Yes	HD-319C	USW G-1	149.9	38C-6	3.7	13.9	00016018
☑ Yes	HD-319C	USW G-1	149.9	38C-6	3.5	13.6	00016018
☑ Yes	HD-319C	USW G-1	149.9	47C-6	3.4	13.3	00016018
☑ Yes	HD-319D	US	149.9	39C-4	2.0	13.7	00016018
☑ Yes	HD-319D	US	149.9	39C-4	2.1	13.7	00016018
☑ Yes	HD-319J	USW G-1	149.9	47C-7	2.1	13.6	00016018
☑ Yes	HD-319E	USW G-1	149.9	38C-8	-1.4	15.3	00016018
☑ Yes	HD-319E	USW G-1	149.9	47C-8	-1.5	15.2	00016018
☑ Yes	HD-320A	USW G-1	176.6	38C-10	-6.8	15.2	00016019
☑ Yes	HD-320A	USW G-1	176.6	38C-10	-6.9	15.1	00016019
☑ Yes	HD-320A	USW G-1	176.6	38C-10	-6.7	15.5	00016019
☑ Yes	HD-320A	USW G-1	176.6	47C-10	-6.8	15.3	00016019
☑ Yes	HD-320B	USW G-1	176.6	39C-1	-6.9	16.4	00016019
☑ Yes	HD-320B	USW G-1	176.6	39C-1	-6.9	16.2	00016019
☑ Yes	HD-320B	USW G-1	176.6	39C-1	-6.8	16.6	00016019
☑ Yes	HD-320B	USW G-1	176.6	45C-8	-6.8	16.5	00016019
☑ Yes	HD-320B	USW G-1	176.6	48C-1	-6.8	16.4	00016019
☑ Yes	HD-320C	USW G-1	176.6	45C-9	-7.7	17.0	00016019
☑ Yes	HD-320C	USW G-1	176.6	48C-2	-8.0	17.0	00016019
☑ Yes	HD-322A	USW G-1	204.0	39C-2	-0.6	14.5	00016021
☑ Yes	HD-322A	USW G-1	204.0	39C-2	-0.7	14.5	00016021
☑ Yes	HD-322A	USW G-1	204.0	39C-2	-0.5	14.7	00016021
☑ Yes	HD-322A	USW G-1	204.0	48C-3	-0.7	14.5	00016021
☑ Yes	HD-322B	USW G-1	204.0	39C-5	-4.9	14.6	00016021
☑ Yes	HD-322B	USW G-1	204.0	48C-4	-5.1	14.5	00016021

Figure E-12 Borehole USW G-1, stable isotope data, calcitic veins.
From D.J.E., 1993.

CALCITE/SILICA - CARBONATE STABLE ISOTOPES

Accept?	ID #	Locality	Depth : m	Estn Log #	d13C pdb	d18O smow	SFC #
☒ Yes	HD-355B	USW G-2	236.7	55C-4	-7.7	17.9	00016055
☒ Yes	HD-356B	USW G-2	240.7	54C-10	-7.9	19.0	00016056
☒ Yes	HD-357A	USW G-2	248.0	55C-2	-8.7	19.0	00016057
☒ Yes	HD-358B	USW G-2	257.8	55C-3	-7.5	17.4	00016058
☒ Yes	HD-359A	USW G-2	178.4	55C-5	-7.7	18.3	00016060
☒ Yes	HD-360A	USW G-2	258.2	55C-6	-8.0	19.3	00016059
☒ Yes	HD-361A	USW G-2	280.2	55C-7	-8.4	19.1	00016061
☒ Yes	HD-362B	USW G-2	280.2	55C-8	-8.3	18.7	00016062
☒ Yes	HD-365A	USW G-2	304.2	56C-1	-7.8	18.1	00016066
☒ Yes	HD-366B	USW G-2	304.2	56C-2	-8.0	18.8	00016066
☒ Yes	HD-366C	USW G-2	304.2	56C-5	-8.0	19.0	00016066
☒ Yes	HD-367A	USW G-2	324.5	56C-7	-6.9	16.7	00016067
☒ Yes	HD-367C	USW G-2	324.5	56C-6	-7.2	17.2	00016067
☒ Yes	HD-368A	USW G-2	346.7	57C-9	-8.1	18.3	00016068
☒ Yes	HD-368B	USW G-2	346.7	57C-10	-7.4	16.8	00016068
☒ Yes	HD-370A	USW G-2	360.8	57C-2	-7.9	19.1	00016070
☒ Yes	HD-370B	USW G-2	360.8	57C-3	-7.9	19.0	00016070
☒ Yes	HD-577A	USW G-2	1254.7	57C-4	-7.6	15.8	00016081
☒ Yes	HD-578A	USW G-2	1256.2	57C-5	-0.4	4.0	00016082
☒ Yes	HD-579A	USW G-2	1268.8	57C-7	-0.2	4.7	00016083
☒ Yes	HD-579B	USW G-2	1268.8	57C-6	-0.1	6.3	00016083
☒ Yes	HD-580A	US	1272.3	57C-8	-0.2	8.8	00016084
☒ Yes	HD-581B	US	1493.6	57C-10	-0.7	6.6	00016085
☒ Yes	HD-582A	USW G-2	1497.4	58C-1	-2.1	6.5	00016086
☒ Yes	HD-582B	USW G-2	1497.4	58C-2	-0.6	6.9	00016086
☒ Yes	HD-583A	USW G-2	1556.7	58C-5	1.1	2.0	00016087
☒ Yes	HD-583A	USW G-2	1556.7	72C-3	1.1	2.0	00016087
☒ Yes	HD-584A	USW G-2	1559.6	58C-6	1.8	4.9	00016088
☒ Yes	HD-585A	USW G-2	1618.0	58C-7	1.2	8.5	00016089
☒ Yes	HD-586A	USW G-2	1624.7	58C-8	0.9	9.9	00016090
☒ Yes	HD-587B	USW G-2	1651.7	58C-10	1.3	8.9	00016091
☒ Yes	HD-588A	USW G-2	1655.4	59C-1	1.2	8.2	00016092
☒ Yes	HD-588A	USW G-2	1655.4	61C-6	1.2	8.4	00016092
☒ Yes	HD-589A	USW G-2	1668.7	59C-2	0.6	9.2	00016093
☒ Yes	HD-590A	USW G-2	1717.8	59C-3	-1.4	5.8	00016094
☒ Yes	HD-590A	USW G-2	1717.8	59C-4	-1.5	5.8	00016094
☒ Yes	HD-591A	USW G-2	1732.2	59C-5	-1.9	8.8	00016095
☒ Yes	HD-591B	USW G-2	1732.2	59C-6	-1.4	9.0	00016095
☒ Yes	HD-593A	USW G-2	1757.1	61C-8	0.0	8.1	00016097
☒ Yes	HD-594A	USW G-2	1760.7	60C-4	-0.2	6.8	00016098
☒ Yes	HD-595A	USW G-2	1763.4	60C-7	-0.8	7.5	00016099
☒ Yes	HD-595A	USW G-2	1763.4	61C-5	-0.8	7.7	00016099
☒ Yes	HD-595C	USW G-2	1763.4	60C-9	-0.3	8.4	00016099
☒ Yes	HD-596A	USW G-2	1786.9	60C-10	0.4	9.9	00016100
☒ Yes	HD-597A	USW G-2	1792.2	61C-2	-0.1	7.3	00016101
☒ Yes	HD-597B	USW G-2	1792.2	61C-3	1.1	10.8	00016101

Figure E-13. Borehole USW G-2, stable isotope data, calcitic veins.
From D.O.E., 1993.

CALCITE/SILICA - CARBONATE STABLE ISOTOPES

Accept?	HD #	Locality	Depth : m	Est. Log #	d13C pdh	d18O smow	SFC #
☒ Yes	HD-261A	USW G-2	1279.9	54C-4	0.8	4.1	00004675
☒ Yes	HD-261A	USW G-2	1279.9	71C-4	-0.1	3.9	00004675
☒ Yes	HD-261B	USW G-2	1279.9	350 C-4	-0.6	7.8	00004675
☒ Yes	HD-262A	USW G-2	1459.4	350 C-7	-1.5	7.8	00004676
☒ Yes	HD-262B	USW G-2	1459.4	350 C-8	-1.0	8.1	00004676
☒ Yes	HD-263A	USW G-2	1636.8	350 C-9	1.1	9.4	00004677
☒ Yes	HD-264A	USW G-2	1639.5	347 C-8	1.9	12.9	00004678
☒ Yes	HD-264B	USW G-2	1639.5	347 C-9	0.7	6.2	00004678
☒ Yes	HD-264C	USW G-2	1639.5	349 C-7	-0.4	8.9	00004678
☒ Yes	HD-265A	USW G-2	1756.3	351 C-1	-0.8	7.1	00004679
☒ Yes	HD-265B	USW G-2	1756.3	351 C-2	-0.4	8.6	00004679
☒ Yes	HD-265C	USW G-2	1756.3	351 C-3	-0.5	9.7	00004679
☒ Yes	HD-266A	USW G-2	1704.1	349 C-9	1.4	10.8	00004680
☒ Yes	HD-272A	USW G-2	441.4	348 C-5	-5.7	14.7	00004686
☒ Yes	HD-272B	USW G-2	441.4	348 C-6	-8.2	15.6	00004686
☒ Yes	HD-272B	USW G-2	441.4	349 C-6	-8.3	15.6	00004686
☒ Yes	HD-272B	USW G-2	441.4	71C-6	-8.4	15.4	00004686
☒ Yes	HD-272D	USW G-2	441.4	348 C-7	-3.6	14.5	00004686
☒ Yes	HD-273A	USW G-2	463.9	348 C-1	-7.9	17.1	00004687
☒ Yes	HD-273B	USW G-2	463.9	348 C-2	-6.0	15.1	00004687
☒ Yes	HD-273C	USW G-2	463.9	348 C-3	-7.9	17.3	00004687
☒ Yes	HD-273D	US	463.9	348 C-4	-4.1	14.7	00004687
☒ Yes	HD-273E	US	463.9	350 C-5	-6.8	15.5	00004687
☒ Yes	HD-273E	USW G-2	463.9	71C-10	-7.0	15.4	00004687
☒ Yes	HD-273F	USW G-2	463.9	350 C-6	-7.3	15.7	00004687
☒ Yes	HD-274A	USW G-2	476.1	348 C-8	-1.5	14.4	00004688
☒ Yes	HD-274B	USW G-2	476.1	348 C-9	-0.9	13.8	00004688
☒ Yes	HD-274C	USW G-2	476.1	348 C-10	-4.7	15.9	00004688
☒ Yes	HD-274D	USW G-2	476.1	349 C-1	-7.5	18.3	00004688
☒ Yes	HD-274E	USW G-2	476.1	349 C-2	-8.2	18.3	00004688
☒ Yes	HD-275B	USW G-2	477.3	350 C-1	-2.0	14.8	00004689
☒ Yes	HD-275C	USW G-2	477.3	350 C-2	-0.2	14.0	00004689
☒ Yes	HD-275D	USW G-2	477.3	53C-8	-1.7	14.1	00004689
☒ Yes	HD-351A	USW G-2	92.2	53C-9	-7.6	20.7	00016051
☒ Yes	HD-351C	USW G-2	92.2	53C-10	-6.8	20.5	00016051
☒ Yes	HD-352A	USW G-2	95.9	54C-1	-5.9	19.8	00016052
☒ Yes	HD-352A	USW G-2	95.9	54C-5	-5.9	19.7	00016052
☒ Yes	HD-352A	USW G-2	95.9	54C-3	-5.9	19.8	00016052
☒ Yes	HD-353A	USW G-2	196.9	54C-2	-8.2	19.5	00016053
☒ Yes	HD-353A	USW G-2	196.9	54C-6	-8.2	19.5	00016053
☒ Yes	HD-353B	USW G-2	196.9	54C-7	-8.4	18.9	00016053
☒ Yes	HD-354A	USW G-2	204.3	54C-8	-7.5	17.2	00016054
☒ Yes	HD-355B	USW G-2	236.7	54C-9	-7.7	17.9	00016055

Figure E-14. Borehole USW G-2, stable isotope data, calcitic veins.
From D.O.E., 1993.

CALCITE/SILICA - CARBONATE STABLE ISOTOPES

Accept?	HD #	Locality	Depth : m	Extn Log #	d13C pdb	d18O smow	SFC #
☒ Yes	HD-267A	USW G-3/GU-3)	313.0	351 C-10	-4.3	18.9	00004681
☒ Yes	HD-267B	USW G-3/GU-3)	313.0	352 C-1	-4.8	19.2	00004681
☒ Yes	HD-268A	USW G-3/GU-3)	1309.7	352 C-4	2.2	10.9	00004682
☒ Yes	HD-268B	USW G-3/GU-3)	1309.7	352 C-5	2.2	10.6	00004682
☒ Yes	HD-268C	USW G-3/GU-3)	1309.7	352 C-6	2.4	11.4	00004682
☒ Yes	HD-268D	USW G-3/GU-3)	1309.7	352 C-7	2.3	9.5	00004682
☒ Yes	HD-269A	USW G-3/GU-3)	1346.0	352 C-8	2.0	11.1	00004683
☒ Yes	HD-269B	USW G-3/GU-3)	1346.0	352 C-9	0.6	10.2	00004683
☒ Yes	HD-270A	USW G-3/GU-3)	1464.0	351 C-6	1.7	7.5	00004684
☒ Yes	HD-276A	USW G-3/GU-3)	294.7	351 C-8	-3.9	18.7	00004690
☒ Yes	HD-276B	USW G-3/GU-3)	294.7	351 C-9	-5.2	19.2	00004690
☒ Yes	HD-277A	USW G-3/GU-3)	296.3	351 C-7	-5.4	19.4	00004691
☒ Yes	HD-277A	USW G-3/GU-3)	296.3	67C-10	-5.4	19.2	00004691
☒ Yes	HD-278A	USW G-3/GU-3)	346.9	357 C-9	-6.4	19.1	00004692
☒ Yes	HD-278B	USW G-3/GU-3)	346.9	357 C-10	-5.7	18.9	00004692
☒ Yes	HD-278C	USW G-3/GU-3)	346.9	358 C-1	-6.2	19.0	00004692
☒ Yes	HD-279A	USW G-3/GU-3)	357.8	352 C-2	-5.9	18.9	00004693
☒ Yes	HD-279B	USW G-3/GU-3)	357.8	352 C-3	-5.7	19.4	00004693
☒ Yes	HD-382A	USW G-3/GU-3)	11.5	65C-10	-5.8	19.6	00016102
☒ Yes	HD-383B	USW G-3/GU-3)	14.6	66C-1	-5.6	20.9	00016103
☒ Yes	HD-383C	USW G-3/GU-3)	14.5	66C-2	-6.4	21.0	00016103
☒ Yes	HD-384A	USW G-3/GU-3)	31.5	66C-3	-6.4	20.7	00016104
☒ Yes	HD-385B	USW G-3/GU-3)	45.6	66C-6	-6.5	21.1	00016105
☒ Yes	HD-386A	USW G-3/GU-3)	50.3	66C-8	-6.4	20.7	00016106
☒ Yes	HD-387A	USW G-3/GU-3)	60.7	66C-9	-5.3	20.7	00016107
☒ Yes	HD-388B	USW G-3/GU-3)	62.5	66C-10	-6.4	20.1	00016108
☒ Yes	HD-388C	USW G-3/GU-3)	62.5	67C-2	-5.5	20.7	00016108
☒ Yes	HD-390A	USW G-3/GU-3)	73.5	67C-4	-6.6	20.3	00016110
☒ Yes	HD-390B	USW G-3/GU-3)	73.5	67C-5	-6.4	20.9	00016110
☒ Yes	HD-390C	USW G-3/GU-3)	73.5	67C-6	-6.8	20.5	00016110
☒ Yes	HD-390D	USW G-3/GU-3)	73.5	67C-7	-4.8	20.0	00016110
☒ Yes	HD-390E	USW G-3/GU-3)	73.5	67C-8	-6.7	20.4	00016110
☒ Yes	HD-391A	USW G-3/GU-3)	75.8	67C-9	-6.1	20.3	00016111
☒ Yes	HD-392A	USW G-3/GU-3)	82.8	68C-2	-6.5	20.2	00016112
☒ Yes	HD-392B	USW G-3/GU-3)	82.8	68C-3	-5.4	20.4	00016112
☒ Yes	HD-392B	USW G-3/GU-3)	82.8	69C-6	-5.5	20.4	00016112
☒ Yes	HD-395A	USW G-3/GU-3)	100.6	68C-6	-6.0	20.3	00016115
☒ Yes	HD-395B	USW G-3/GU-3)	100.6	68C-7	-5.6	19.9	00016115
☒ Yes	HD-397A	USW G-3/GU-3)	112.1	68C-10	-5.4	20.7	00016117
☒ Yes	HD-397B	USW G-3/GU-3)	112.1	69C-1	-6.6	19.6	00016117
☒ Yes	HD-398A	USW G-3/GU-3)	131.4	69C-2	-5.3	19.9	00016118
☒ Yes	HD-400B	USW G-3/GU-3)	146.2	69C-5	-5.4	20.0	00016120
☒ Yes	HD	USW G-3/GU-3)	151.9	69C-7	-4.2	20.0	00016121

Figure E-15. Borehole USW G-3 / GU-3, stable isotope data, calcitic veins.
From D.O.E., 1993.

CALCITE/VEIN - CARBONATE STABLE ISOTOPES

Accept?	ID #	Locality	Depth : m	Estm Loc #	d13C pdb	d18O smow	SPC #
☒ Yes	HD-403A	USW G-3/GU-3)	168.6	69C-8	-4.1	19.3	00016123
☒ Yes	HD-406A	USW G-3/GU-3)	192.0	69C-9	-5.8	20.0	00016126
☒ Yes	HD-421A	USW G-3/GU-3)	341.7	70C-1	-5.3	18.4	00016141
☒ Yes	HD-422A	USW G-3/GU-3)	349.8	71C-2	-6.0	18.1	00016142
☒ Yes	HD-422/	USW G-3/GU-3)	349.6	70C-2	-6.0	18.2	00016142
☒ Yes	HD-424A	USW G-3/GU-3)	521.4	70C-3	-5.7	15.8	00016144
☒ Yes	HD-598A	USW G-3/GU-3)	893.6	71C-5	2.1	9.6	00016157
☒ Yes	HD-599A	USW G-3/GU-3)	897.9	72C-1	2.4	9.1	00016158
☒ Yes	HD-599B	USW G-3/GU-3)	897.9	72C-4	3.2	8.9	00016158
☒ Yes	HD-600A	USW G-3/GU-3)	900.3	72C-6	3.3	9.0	00016159
☒ Yes	HD-601A	USW G-3/GU-3)	903.6	72C-7	2.2	9.2	00016160
☒ Yes	HD-602B	USW G-3/GU-3)	908.5	72C-9	2.7	9.2	00016161
☒ Yes	HD-606A	USW G-3/GU-3)	938.5	73C-1	3.7	9.4	00016165
☒ Yes	HD-606B	USW G-3/GU-3)	938.5	73C-2	2.0	9.5	00016165
☒ Yes	HD-608A	USW G-3/GU-3)	946.9	73C-4	3.2	9.0	00016167
☒ Yes	HD-610A	USW G-3/GU-3)	1315.6	73C-5	2.3	7.1	00016169
☒ Yes	HD-610B	USW G-3/GU-3)	1315.6	73C-6	2.3	11.0	00016169
☒ Yes	HD-610C	USW G-3/GU-3)	1315.6	73C-7	2.6	7.7	00016169
☒ Yes	HD-611A	USW G-3/GU-3)	1324.8	73C-8	1.6	10.0	00016170
☒ Yes	HD-612A	USW G-3/GU-3)	1330.5	73C-9	1.0	8.3	00016171
☒ Yes	HD-613A	USW G-3/GU-3)	1343.0	73C-10	1.4	11.0	00016172
☒ Yes	HD-614A	USW G-3/GU-3)	1349.5	74C-2	2.1	11.0	00016173
☒ Yes	HD-614B	USW G-3/GU-3)	1349.5	74C-3	1.9	11.2	00016173
☒ Yes	HD-615A	USW G-3/GU-3)	1407.9	74C-4	1.6	7.8	00016174
☒ Yes	HD-617A	USW G-3/GU-3)	1456.2	74C-6	0.9	12.0	00016176
☒ Yes	HD-619A	USW G-3/GU-3)	1521.0	74C-8	2.3	10.6	00016178
☒ Yes	HD-620A	USW G-3/GU-3)	1530.0	74C-10	0.9	8.9	00016179
☒ Yes	HD-722A	USW G-3/GU-3)	13.3	70C-4	-6.6	20.6	00005520
☒ Yes	HD-723A	USW G-3/GU-3)	14.4	70C-7	-6.6	20.7	00005517
☒ Yes	HD-723B	USW G-3/GU-3)	14.4	70C-8	-6.1	20.9	00005519
☒ Yes	HD-724A	USW G-3/GU-3)	17.3	70C-9	-6.6	20.9	00005518
☒ Yes	HD-725A	USW G-3/GU-3)	22.1	70C-10	-6.4	20.2	00005517
☒ Yes	HD-726A	USW G-3/GU-3)	27.4	71C-1	-6.3	20.4	00005516

Figure E-16. Borehole USW G-3 / GU-3, stable isotope data, calcitic veins.
From D.O.E., 1993.

CALCITE/SILICA - CARBONATE STABLE ISOTOPES

Accept?	ID #	Locality	Depth : m	Extm Log #	d13C pdb	d18O smow	SPC #
☒ Yes	HD-926B	UE-25 A-5	85.2	63C-9	-6.3	17.8	00019106
☒ Yes	HD-926C	UE-25 A-5	85.2	63C-10	-6.4	17.7	00019106
☒ Yes	HD-927A	UE-25 A-5	85.9	64C-1	-2.8	14.7	00019107
☒ Yes	HD-927B	UE-25 A-5	85.9	64C-2	-6.7	17.8	00019107
☒ Yes	HD-927C	UE-25 A-5	85.9	64C-4	-6.4	17.7	00019107
☒ Yes	HD-928A	UE-25 A-5	86.9	64C-5	-7.5	18.6	00019108
☒ Yes	HD-928B	UE-25 A-5	85.9	64C-6	-6.5	18.3	00019108
☒ Yes	HD-929B	UE-25 A-5	92.2	64C-7	-5.8	17.9	00019109
☒ Yes	HD-931A	UE-25 A-5	104.8	64C-8	-6.8	17.8	00019111
☒ Yes	HD-932A	UE-25 A-5	108.5	64C-10	-5.2	15.9	00019112
☒ Yes	HD-933A	UE-25 A-5	110.9	65C-1	-6.2	17.6	00019113
☒ Yes	HD-935A	UE-25 A-5	140.2	65C-3	-5.9	17.3	00019115
☒ Yes	HD-936A	UE-25 A-5	143.2	66C-5	-6.2	17.5	00019116
☒ Yes	HD-937A	UE-25 A-5	145.8	66C-6	-6.0	15.2	00019117
☒ Yes	HD-937B	UE-25 A-5	145.8	65C-7	-6.0	16.2	00019117
☒ Yes	HD-938B	UE-25 A-5	147.7	65C-9	-6.1	16.8	00019118

**Figure E-17. Borehole UE-25 A-5, stable isotope data, calcitic veins.
From D.O.E., 1993.**

Sample #	Drill Hole	Depth (m)	Type ^a	$\delta^{13}C^b$	$\delta^{18}O^c$
HD-230	G-1	185	vein	-5.5	14.2
HD-231	G-1	195	vein	-1.3	11.0
	G-2 ^d	180		-4.4	19.2
	G-2 ^d	102		-7.9	19.3
	G-2 ^d	117		-7.4	18.2
	G-2 ^d	147		-7.4	18.3
	G-2 ^d	110		-7.5	18.2
	G-2 ^d	119		-6.9	18.1
	G-2 ^d	159		-6.8	18.0
HD-272A	G-2	111	vug	-5.7	14.7
HD-272B	G-2		vug	-8.3, -8.3	15.6, 15.6
HD-272D	G-2		vein	-3.6	14.5
HD-273A	G-2	164	vug 1	-4.2	14.7
HD-273B	G-2		vug 2	-7.9	17.3
HD-273C	G-2		vug 3	-6.0	15.1
HD-273D	G-2		vug 4	-7.9	17.1
HD-273E	G-2		vein 1	-6.9	15.5
HD-273F	G-2		vein 2	-7.3	15.7
HD-274A	G-2	176	vug 1	-1.5	14.4
HD-274B	G-2		vug 2	-0.9	13.8
HD-274C	G-2		vug 3	-4.7	15.9
HD-274D	G-2		vug 4	-7.6	18.3
HD-274E	G-2		vug 5	-8.2	18.3
HD-275B	G-2	177	vug 2	-2.0	14.8
HD-275C	G-2		vug 3	-0.3	14.0
HD-261A	G-2	1280	vein	0.0	4.2
HD-261B	G-2		cement	-0.6	7.8
HD-262A	G-2	1459	vein	-1.5	7.8
HD-262B	G-2		cement	-1.0	8.1
HD-263A	G-2	1637	vein	1.0	9.4
HD-263B	G-2		cement	0.4	7.8
HD-264A	G-2	1640	vein 1	1.9	12.9
HD-264B	G-2		vein 2	0.7	6.2
HD-264C	G-2		cement	-0.4	8.9
HD-265A	G-2	1756	vein 1	-0.4	8.6
HD-265B	G-2		vein 2	-0.8	7.1
HD-265C	G-2		cement	-0.5	9.7
HD-266A	G-2	1794	cement	1.3	10.5
	G-3 ^d	63		-7.1	20.2
	G-3 ^d	131		-5.1	20.2
	G-3 ^d	147		-5.6	20.0
	G-3 ^d	159		-5.4	20.3
HD-276A	GU-3	295	vein?	-4.0	18.7

Explanation:

- a) numbered sequences are from early to late.
- b) $\delta^{13}C$ vs. PDB.
- c) $\delta^{18}O$ vs. SMOW.
- d) analyses reported by Szabo and Kyser⁴.
- e) unpublished data from B.A. Carlos.

Figure E-18. Isotopic character of carbon and oxygen incorporated in the Yucca Mountain veins. From Whelan and Stuckless, 1991.

Sample #	Drill Hole	Depth (m)	Type	$\delta^{13}C$	$\delta^{18}O$
HD-276B	GU-3	-	vein?	-5.2	19.2
HD-277A	GU-3	196	vein	-5.4	19.4
HD-267A	G-3	313	vein 1	-4.3	18.9
HD-267B	G-3	-	vein 2	-4.8	19.2
	G-3 ^d	318		-5.1	19.1
	G-3 ^d	331		-4.5	18.7
HD-278A	GU-3	347	vein 1	-5.7	18.9
HD-278B	GU-3	-	vein 2	-6.3	19.0
HD-278C	GU-3	-	veinlet 3	-6.5	19.1
HD-279A	GU-3	358	vein 1	-6.0	18.9
HD-279B	GU-3	-	vein 2	-5.7	19.4
HD-268A	G-3	1310	vein	2.2	10.9
HD-268B	G-3	-	vein	2.1	10.6
HD-268C	G-3	-	vein	2.4	11.3
HD-268D	G-3	-	vein	2.3	9.5
HD-269A	G-3	1346	vein 1	0.5	9.7
HD-269B	G-3	-	vein 2	-2.0	11.7
HD-270A	G-3	1464	vein	1.7	7.5
HD-255A	b1	942	vein	3.8	5.3
HD-256A	b1	953	vein A	2.4	5.8
HD-256B	b1	-	vein B	1.8	5.8
HD-257A	b1	971	vein 1	4.2	11.7
HD-257B	b1	-	vein 2	4.4	11.4
HD-258A	b1	1098	vein	2.0	6.0
HD-259A	b1	1116	vein 1	1.6	5.7
HD-259B	b1	-	vein 2	0.6	4.4
HD-259C	b1	-	cement	2.1	6.8
HD-260A	b1	1216	vein 1	1.3	5.5
HD-260B	b1	-	vein 2	2.1	4.6
HD-260A	b1	-	cement	1.0	8.8
	G-4 ^e	77		-6.8	17.0
HD-674A	G-4	108	vug 1	4.9	16.0
HD-674B	G-4	-	vug 2	0.6	16.5
HD-679A	G-4	265	vug 1	-7.1	16.8
HD-679B	G-4	-	vug 2	-1.2	14.7
	G-4 ^e	-		-0.8	14.1
	G-4 ^e	353		-3.6	17.1
HD-681A	G-4	401	vein	-9.1	16.0
	G-4 ^e	-		-9.0	15.2
	G-4 ^e	587		-6.7	17.4
	G-4 ^e	663		-7.9	17.6
HD-687A	G-4	810	vein	-8.2	13.6
	G-4 ^e	-		-8.0	13.1
HD-689A	G-4	842	vein	-5.7	13.3
	G-4 ^e	-		-6.4	13.2

Explanation:

- numbered sequences are from early to late.
- $\delta^{13}C$ vs. PDB.
- $\delta^{18}O$ vs. SMOW.
- analyses reported by Szabo and Kyser⁶.
- unpublished data from B.A. Carlos.

Figure E-19. Isotopic character of carbon and oxygen incorporated in the Yucca Mountain veins. From Whelan and Stuckless, 1991.

Sample	Material dated	U-series age [$\times 10^3$ yrs B.P.]	Source
Borehole USW G-2			
280	calcite	>400	Szabo and Kyser (1985)
346.7	do	190 ± 20	do
348.7	do	142 ± 30	do
348.8	do	>400	do
348.8	do	280 ± 70	do
359-A	do	170 ± 18	do
359-B	do	185 ± 18	do
Borehole USW GU-3			
63	calcite	227 ± 20	Szabo and Kyser (1985)
131	do	26 ± 2	do
318	do	>400	do
331	do	30 ± 4	do
Borehole UE-25a#1			
34	calcite	$310 \pm \begin{smallmatrix} 60 \\ 50 \end{smallmatrix}$	Szabo and Kyser (1985)
283	do	$310 \pm \begin{smallmatrix} 70 \\ 45 \end{smallmatrix}$	do
611	do	>400	do

Figure E-20. U-series ages, samples of the Yucca Mountain veins.

Figure E-21. U-series ages from sample of the Yucca Mountain calcretes, surficial calcitic veins, and subsurface calcitic veins.

APPENDIX F

APPENDIX F

"Thermal Ground-Water Movement and Radionuclide Transport in SW England,"

by E. M. Durrance and M. J. Heath

"Modeling of Saturated Zone at Yucca Mountain, Nevada,"

by M. P. Ahola and B. Sager

Thermal groundwater movement and radionuclide transport in SW England

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ABSTRACT. Heat flow in SW England is well above average for the UK as a whole, but northwards towards Bath and Bristol the values decrease rapidly. However, hot springs occur both in the Bath-Bristol area and in mines in Cornwall. The development of hydrothermal circulation systems is thus not controlled entirely by geothermal gradient: the presence of a suitable fracture permeability is the main requirement. The thermal 'head' produced, which nevertheless depends upon the temperature and volume of water in the system, theoretically can exceed local topographic heads even in areas of low geothermal gradient.

Thermal groundwaters usually carry above average quantities of radionuclides in solution because of the long residence times involved. ^{222}Rn values are often particularly high. High concentrations of ^{222}Rn in surface waters arise from the discharge of groundwater. The results of a survey of ^{222}Rn in streams in SW England have established areas of high values which are interpreted as rising limbs of convection cells with dimensions in the order of 5-10 km.

In SE Devon γ -ray spectrometry of soils shows two E-W belts of high activity. The northern is coincident with the faulted southern margin of the Crediton Trough, while the southern is coincident with the westerly extension of the Abbotsbury fault system. Groundwater movement along deep-seated fracture systems is considered to be the explanation of these features. The horizontal scale of the area involved suggests that a thermal rather than local topographic head is the driving force. Groundwater circulation within fractures, driven by a thermal head, may therefore occur even in areas of low geothermal gradient and should be considered when selecting waste disposal sites.

KEYWORDS: radionuclide transport, groundwaters, SW England, radon.

THE disposal of radioactive waste by deep burial in crystalline rock such as granite, has been one of the options considered during the development of a radioactive waste management policy for the UK. Field investigations at an experimental site in the Carnmenellis granite in Cornwall have been concerned principally with fracture permeability (Heath, 1985), the results of these studies being used as part of the basis for modelling water movement

through the rock under conditions of low, topographically induced hydraulic head. However, as the area of SW England underlain by granite is characterized by heat flow values considerably higher than the average for the British Isles, and by the products of a long history of hydrothermal activity, notably metalliferous mineralization as late as Tertiary, kaolinization, and the occurrence of hot springs in some Cornish mines, the possibility of an additional thermal head as an influence on modern groundwater movement, and, therefore, on radionuclide migration, should also be considered.

Although heat flow values in SW England fall rapidly away from the granite, in the Bath-Bristol area, where they are actually well below average, hot springs occur, suggesting that the influence of thermally induced head on groundwater movement may extend to areas of average heat flow where it is probably controlled by the presence of major fractures. If geological isolation of radioactive waste is in sedimentary rather than granite rocks the possibility of thermal groundwater movement, even in areas of average heat flow, should therefore still be considered.

A difficulty in assessing the importance of present day thermal groundwater movement is that evidence of such activity may only reflect past circulation patterns and not a modern dynamic system. An investigation of this problem has therefore been made, based upon the mobility of U and some of its decay products, and other naturally radioactive elements in aqueous solutions, so that circulation systems may be first identified and secondly recognized as either fossil or active.

Hydrothermal circulation

Geothermal Systems may be divided into two broad categories: Convective Geothermal Systems and Conductive Geothermal Systems, depending upon the dominant nature of the heat transfer process. Convective geothermal systems, in turn,

may be subdivided into Hydrothermal and Circulation Systems, while conductive systems are subdivided into Low Temperature and Hot Dry Rock systems (Rybach, 1981).

The characteristic environments in which each of these systems occurs is relatively straightforward. Hydrothermal systems are found in areas where the rocks have high porosity and permeability and are related to shallow, young intrusions. Circulation systems occur in low porosity/fracture permeability rocks where the regional heat flow is generally above normal. Low-temperature conductive systems are found in high porosity/permeability sedimentary rocks where the heat flow is usually above normal and hot dry rock systems occur in high temperature but low permeability environments.

Convective geothermal systems manifest themselves at the surface of the Earth by fumaroles and boiling springs, and so are very apparent. Circulation systems, however, are less obvious, but discharge may be characterized by hot springs, particularly when related to faults or fracture zones in topographically low ground. Recharge may also occur along such fracture zones. The general crustal setting for circulation systems is usually recent extensional tectonism especially where this is accompanied by a high regional heat flow. In these environments fault and fracture zones are often kept open by periodic seismic activity, and basement fractures are particularly important in controlling the location of rising groundwater. Within these zones spring discharge temperatures are often strongly influenced by the admixture of cold surface waters to the ascending thermal water.

Heat flow in southern Britain. Much attention has recently been focused on SW England as a potential source of geothermal energy (Garnish, 1976). Heat flow studies by Oxburgh *et al.* (1977) and Richardson and Oxburgh (1979) have shown that the average heat flow value for the UK is about 1.43 hfu (heat flow units or $10^{-6} \text{ cal cm}^{-2} \text{ s}^{-1}$), corresponding to a geothermal gradient of 25°C/km which is comparable with the average value of about 1.53 hfu for continental Europe (Cermak, 1979). In a linear belt stretching approximately E.-W. from the Isles of Scilly to near Southampton, however, heat flow values are greater than 1.91 hfu and reach values greater than 2.39 hfu (associated with a geothermal gradient of up to 40°C/km) in the zone above the SW England granite batholith (Haenal, 1980).

The high heat flow of SW England was believed by Richardson and Oxburgh (1979) to be due to the presence of granite bodies within the upper crust, which carry above average concentrations of heat-producing radioactive elements. They therefore

envisaged essentially a hot dry rock system within the granite areas, but considered that limited circulation systems could locally modify the heat flow distribution.

Rock permeabilities. The ability for mass transfer to take place within rock is determined by both the intergranular permeability of highly porous rocks and the presence of fractures in rocks which can have a great range of inherent porosities. Fluid flow in fissured rocks differs in several important respects, however, from the more pervasive percolation that takes place in competent porous rocks. Garg and Kassoy (1981) listed three main areas of difference:

(1) Fracture-induced permeability is usually much greater than permeability due to interconnected pore spaces. Louis (1970), in fact, noted that the matrix permeability of rock becomes important only in the absence of continuous joints or for joint apertures less than $10 \mu\text{m}$.

(2) Fracture permeability is usually anisotropic.

(3) Fracture porosity (and hence permeability) is much more sensitive to fluid pressure and rock stresses than matrix porosity (Wittke, 1973; Duncan *et al.*, 1972).

The spacing of fractures is an important factor when describing the characteristics of flow. For if the fracture spacing is close in terms of the total linear dimensions of the system considered, then processes which are dominantly fracture controlled on a small scale may be better explained in terms of pervasive flow when considered on a large scale.

In terms of flow in single fractures, the pipe flow model of Donaldson (1968) is most appropriate. With direct connection down from the surface, the thermal hydraulic heads produced may be very large. For example, Elder (1981) has calculated that in the Wairakei geothermal field (New Zealand), where the temperature is about 250°C at a depth of about 5 km, the equivalent head produced is as great as 1 km.

In established geothermal fields Elder (1981) has also noted that a permeability of 0.1 d is more than adequate to permit all the hydrothermal phenomena observed, but that the contribution of the intergranular permeability to this is negligible. He made the interesting order of magnitude calculation of the permeability afforded by fractures, by considering a system in which the rock is a cubic array of blocks of width Δ_1 , with narrow spaces of width Δ_2 separating them. Then if $\Delta_1 = 1 \text{ m}$ and $\Delta_2 = 0.1 \text{ mm}$ a permeability of 0.2 d results, but if $\Delta_1 = 100 \text{ m}$ and $\Delta_2 = 0.5 \text{ mm}$ we still get 0.2 d, and 0.2 d is also found if $\Delta_1 = 1000 \text{ m}$ and $\Delta_2 = 1 \text{ mm}$. It is thus easy to see that the occurrence of the occasional fracture (which need not even be large)

gives a very significant permeability to the rock mass as a whole.

Kasoy and Zehib (1978) also considered the cooling of a rising column of water in a vertical channel of porous material with impermeable walls, and found that typical vertical convective rates were about 6×10^{-3} m/d. Even at this speed the loss of heat by conduction from the fault zone was very little compared with the convective heat transfer.

Fault controlled charging of a geothermal reservoir has been studied by Goyal and Kasoy (1977) and Goyal (1978). The model they chose considered a fracture zone which extended down through a surface impermeable layer, then through an interbedded permeable and impermeable series of layers, ending finally in a layer of basement rock. As in the Kasoy and Zehib (1978) model, the fault zone was taken to be charged with hot water flowing from the fractured basement rock. In this model it was found that hot water rose through the fracture, as before, but the presence of the surface impermeable cap suppressed vertical transport so that water was pushed out of the fault zone by the thermal head and into the permeable horizons in the interbedded series. Horizontal flow in these permeable horizons then became dominant. It was found that in this system horizontal distances away from the fault zone of 5-10 fault depths had to be traversed before the aquifers ceased to be affected by the outflow from the fault zone. This clearly shows that fault zones can provide a mechanism for charging shallow geothermal aquifers.

Convective circulation of water in the granite of South West England

Framework for convective groundwater circulation. In considering the possible existence of convective water circulation it must first be established that the physical requirements for such circulation are satisfied and that circulation is theoretically possible, given the properties of the crust in that region. Secondly, the effects of such circulation should be investigated, particularly in terms of the transport of heat and certain mobile elements as reflected in their distribution at the surface. Such a twofold approach has been adopted in the area underlain by the Dartmoor granite and it appears that modern hydrothermal circulation may indeed be of significance in this and the other granite plutons of the region.

The requirements for convective water circulation in the crust are, first, that a source of heat is available; secondly, that there is sufficient permeability within the required vertical range, and thirdly that groundwater is present. That a heat

source is available has been established (Wheildon *et al.*, 1980). The anomalously high heat flow of the area underlain by granite in SW England (up to 3.0 hfu) is believed to be due largely to the heat generated by the decay of radioelements in the granite (notably U). Although fabric permeabilities for the granite of SW England are as low as 100 pd (Batchelor, 1980), the highly fractured nature of the granite produces an *in situ* permeability many orders of magnitude higher. Thus, the hydrogeological investigations in a 700 m borehole in the Carnmenellis granite (Heath, 1985) have shown that major water-bearing fractures exist at depths of at least 650 m. Indeed, Alderton and Sheppard (1977) reported hot springs issuing from depths of up to 900 m below Ordnance Datum in certain Cornish mines. Many of these thermal waters have high salinities (up to 1.5% by weight total dissolved solids) although their hydrogen and oxygen isotopic compositions indicate a meteoric origin. The high salinities therefore result from ion-exchange processes in the aquifer system and the presence of deep groundwater in the granite is indicated.

Hydrogeochemistry and convective groundwater movement in the Dartmoor granite. A hydrogeochemical survey was carried out in the Dartmoor area in connection with an investigation into the distribution of U in the granite (Heath, 1982). Surface waters from 538 localities (391 within the area underlain by granite) were analysed for ^{222}Rn by scintillation counting following degassing of radon, while 222 of these samples (185 from within the area underlain by granite) were analysed for U by laser-induced fluorimetry.

The observed distribution of Rn and U in Dartmoor stream waters was found not to be related entirely to the known distribution of U in the underlying rocks. Rn distribution was shown to be controlled largely by hydrological factors unrelated to the U content of the bedrock, and many of the well-developed Rn anomalies may be attributed to the influx of groundwater (Mogro-Campero and Fleischer, 1977) in a pattern not related to topography.

Of most significance is the observation that the high Rn concentrations of the streams of central Dartmoor (Vitifer) occur in a zone where particularly high permeability of the granite is indicated by electrical resistivity work (Durrance *et al.*, 1982), and in an area of very high heat flow (Wheildon *et al.*, 1980). This suggests the possibility that the high Rn levels reflect water movement in a deep fracture system, with upwelling by thermal convection. Similar upflows appear to occur in the area south of Okehampton, in NE Dartmoor and in S. Dartmoor.

If the areas of convective ascent are associated with transport of Rn and U to the surface it is likely

that areas of convective drawdown are characterized by a surface environment depleted in these elements. It is of interest, therefore, that the area between 5 and 10 km north of the Rn anomaly around Vitifer is characterized by low stream Rn and U concentrations and is also traversed by the Sticklepath fault zone (Dearman, 1963). It is possible that this represents the descending limb of the convection cell, the ascending limb of which occurs around Vitifer. Similar drawdown may be taking place along a topographically-inferred fracture zone west of Vitifer with corresponding ascent south of Okehampton. Thus, in northern Dartmoor at least two convection cells may be postulated with dimensions of approximately 5 km.

In SW Dartmoor the area in which the granite is known to be kaolinized is also typified by generally low stream Rn and U concentrations. This is consistent with the known leaching of U from the granite during kaolinization. However, in the light of the evidence from N. Dartmoor this association may be more significant. The kaolin deposits of SW England usually occur in upward-facing funnel-shaped structures which, although possibly initiated by partial alteration soon after granite emplacement, are related largely to the effects of meteoric waters (Bristow, 1977; Sheppard, 1977). The association of kaolinization with major faulting in SW Dartmoor (the Calisham Down-Cornwood fault, plus many others inferred on topographical grounds) is significant in this connection. It is suggested, therefore, that this area is the site of convective drawdown in a cell the ascending limb of which is probably reflected in the high Rn concentrations of the streams immediately north of the kaolinized granite.

A summary map of Dartmoor, showing some of the evidence on which this interpretation is based, is given in fig. 1. Three areas of hydrothermal discharge (ascent) and three associated areas of recharge (descent) are indicated. Using flow data from Fehn *et al.* (1978) and assuming an average *in situ* permeability of 100 μ d and cell dimensions of 5 km, the residence time of circulating water below the surface of the granite can be estimated as 10^3 – 10^6 years. Evidence from Cornish mines regarding the flow of thermal waters from cross-courses and from the electrical resistivity survey of Dartmoor (Durrance *et al.*, 1982) suggests that the main orientation of flow paths in these hydrothermal systems is N.–S. to N.–NW–S.–SE; that is approximately parallel to the maximum horizontal stress which tends to keep E.–W. fractures closed.

In plutons of uniform permeability convection cells may migrate with time but, where the granite is structurally inhomogeneous, systems become anchored to major structural features and remain

in the same location for long periods of time (Mogro-Campero and Fleischer, 1977). Assuming that the major requirements for convective circulation (a heat source, sufficient permeability and the presence of groundwater) have been satisfied, the convective systems postulated from Dartmoor may have been active for a considerable period of geological time.

Thermal groundwater movement in SE Devon

Hydrogeochemical evidence. High values of stream water ^{222}Rn occur near the mouth of the Exe Estuary (from samples collected on both sides of the valley) and in a zone which extends northwest to beyond Kennford (Durrance, 1978). The anomaly thus trends diagonally across the New Red Sandstone outcrop from the Littleham Mudstone Formation to the Kennford Breccias (fig. 2) and is parallel to the set of NW–SE trending transcurrent faults which affect SW England (Dearman, 1963). As for Dartmoor, it is concluded that the most likely mechanism that will allow the formation of the observed ^{222}Rn anomaly is the discharge of groundwater. The groundwater discharge, in turn, is considered to be controlled by the presence of a NW–SE fracture zone passing through Kennford towards Exmouth. Here, however, the driving force for the groundwater flow could be either a topographic head or a geothermal head, and although modern groundwater flow must be taking place to account for the presence of the radon anomaly, it cannot be determined for how long this has happened.

γ -ray surveys. An airborne γ -ray survey of SW England which extended as far east as Budleigh Salterton was carried out for the Geological Survey in the late 1950s. The main centres of high activity that were detected occur in three geographical positions: a narrow N.–S. trending zone in the Teign Valley (probably related to a U–Th mineral vein), a broad area around Kennford, the area south of Crediton and a belt trending north from Littleham Cove. Many minor, local high-activity anomalies were also discovered. Systematic interpretation of these results was difficult because of the problems caused by radioactive fall-out from the atmospheric testing of nuclear weapons that was taking place at the time of the survey. It is considered that many of the minor occurrences of high activity have their origin in radioactive fall-out. The main zones of high activity, however, occur in positions which correlate with the results of a stream water U survey (Durrance, 1984), where three belts of high activity trend N.–S. The most westerly of these is related to the breccias, the central to the Littleham Mudstone Formation, and

FIG. 1. Summary map of the Dartmoor granite showing possible distribution of Convective Circulation Systems.

FIG. 2. Geology of SE Devon.

the eastern to the lower part of the Mercia Mudstone Formation.

Ground-based γ -ray spectrometry of soils at 559 sites in SE Devon (Durrance, 1983), using a Geometrics DISA 400A and a count-time of five minutes at each site (Lovborg *et al.*, 1971; Lovborg, 1972; Cassidy, 1982), also shows that the main areas of high activity recorded by the airborne survey are not the results of short-lived fall-out (fig. 3A). However, the area north of Littleham Cove is not characterized by soils with high activity, although clearly having U enrichment in the form of the concretionary nodules. The discrepancy is explained as a product of the different sampling techniques used in the various surveys, and their

interaction with the sporadic nature of the occurrence of the U-bearing nodules. Thus, whereas the airborne and stream water U surveys considered samples which 'averaged' effects from a wide area, because the DISA 400A detector was placed directly on the ground during counting, the soil survey considered only a small volume of material at each site, so the chance of finding any influence of a U-bearing nodule was small. The main additional feature shown by the distribution of the Total Count values in the results of the soil survey is the presence of easterly trending zones of high activity. One zone passes close to Feniton and continues east of Honiton; another, more complex, zone is in the Exmouth-Sidmouth area.

FIG. 3. SE Devon: (A) Total γ -ray activity of soils; (B) U content of soils; (C) K content of soils; (D) Th content of soils.

The U content of the soils, shown by the activity of ^{214}Bi in the γ -ray spectrometer survey, is also high in the easterly trending zones of high Total Count (fig. 3B). The extent of the zone passing through Feniton is shown by the U values to be particularly impressive, the line continuing from Exeter to north of Axminster. High U values also occur in the Teign Valley (where they are related to the mineral vein detected by the airborne survey), and in the NW-SE trending zone through Kennford. U values are not high in the soils over the Littleham Mudstones or the Exmouth Sandstones and Mudstones, again probably reflecting the sporadic distribution of the U-bearing nodules. The lower part of the Mercia Mudstones, however, shows soils in which the U content is consistently above average. This suggests that U-bearing

nodules which occur in these strata are much more widely distributed than in the Permian formations. U levels in the soils on the Tertiary flint gravels and the Cretaceous strata are low. They are also low in the soils which overlie the Bovey Formation.

The total γ -ray activity and U distribution in the soils of SE Devon thus generally confirm the picture given by the results of the stream water U survey, but additionally provide evidence for the presence of approximately E.-W. lines of U enrichment. These lines probably indicate the positions of deep-seated fractures which act as channelways for moving groundwater.

K distribution in the soils of SE Devon (shown by the activity of ^{40}K in the γ -ray spectrometry survey) closely follows that of U and total γ -ray activity, except that the mineral vein in the Teign Valley

does not appear as a K high (fig. 3C). The main occurrences of high K values are found in the area around Kennford, with extensions to the NW and SE, in the E.-W. zone passing through Feniton and in the Exmouth-Sidmouth area. K enrichment in these zones is probably the result of upflow of groundwater carrying K as well as U, in solution. Both U and K are highly mobile elements in aqueous solution and their association in low-moderate temperature hydrothermal systems is expected from their geochemical characteristics.

Th is generally immobile in low moderate-temperature aqueous environments, and the distribution of Th in the soils of SE Devon (shown by the activity of ^{208}Tl in the γ -ray spectrometer survey) reflects the levels of Th in the underlying rocks (fig. 3D). With the exception of Th occurrences in the Teign Valley, the highest Th concentrations are found in association with the mudstone formations in the New Red Sandstone. The N.-S. mineral vein in the Teign Valley, however, is shown by soils with very high Th contents indeed. An association of high U and high Th with low or normal K levels in this feature suggests that the vein carries pitchblende. It is thus similar to other U mineralization in SW England, where N.-S. veins (cross-courses) are often of Tertiary age.

The general distribution of U, K, and Th thus probably results from the original concentrations of these elements in the detrital minerals of the sedimentary succession in SE Devon, but upon this there appears to have been superimposed the effects of mobilization, transport, and precipitation of elements derived from outside the succession. The character of these secondary enrichments varies according to the geochemical properties of the elements concerned, the temperature and chemistry of the mobilizing system and the nature of the host rocks. Thus, Th appears to have been mobilized only in the formation of the Teign Valley mineral vein, but U and K have a more extensive spatial record of mobilization, which probably also reflects a more extensive temporal record of mobilization.

Disequilibrium in the ^{238}U Decay Series. Secular equilibrium within any decay series is achieved by the growth of abundance of the various daughter products in a closed chemical system. This growth is exponential with respect to time so that, theoretically, equilibrium is reached only after an infinite period of time. In practice, therefore, equilibrium is assumed at the 95 or 99% level. Decay of ^{40}K is simple and not liable to disequilibrium, and the ^{232}Th decay series, although long and composed of many daughter elements, is characterized by short half-lives, so that disequilibrium is unlikely. The ^{238}U decay series, however, also has a long and complex decay chain, and some of the

daughter products have long half-lives. It is these circumstances which make the ^{238}U decay series very susceptible to disequilibrium when taken in conjunction with the chemical mobility of the parent and some of the daughters (Adams and Gasparini, 1970).

For the ^{238}U decay series, 99% secular equilibrium is reached after about 1.66 Ma. Thus if disequilibrium in the decay series, resulting from the gain or loss of any daughter (or the parent) in an open chemical system, is found, it implies that these changes have occurred in the recent past, because of the short (geologically speaking) time needed to establish equilibrium. Indeed, if substantial disequilibrium is present, it is the result of processes that have probably operated in about the last 2.5 ka, and may, therefore, still be acting today.

A Disequilibrium Factor has been established, using the results of the γ -ray spectrometer survey of the rocks and soils of SE Devon, in which comparison is made between the Total Count recorded at a particular site, and the count expected from the K, U, and Th concentrations at the site with the ^{238}U decay series in equilibrium (Durrance, 1983). Because the Total Count additionally records the essentially low energy (0.1-1.25 MeV) part of the γ -ray spectrum not covered by the separate channel windows, the main contributor in the ^{238}U decay series is ^{214}Th , but this will be in equilibrium with ^{238}U . Thus where the Total Count is less than expected (negative Disequilibrium Factor) loss of U or addition of Rn has occurred in the soil, and, conversely, where the Total Count is greater than expected (positive Disequilibrium Factor) there has been recent U addition or Rn loss. The pattern of Rn loss from the soils in SE Devon, however, shows no association with the distribution of positive and negative values of the Disequilibrium Factor and it is therefore concluded that the Disequilibrium Factor indicates U mobility (Ostrikhansky, 1976). Negative values of the Disequilibrium Factor dominate the results from SE Devon, showing loss of U by modern leaching to be widespread, but positive values occur in a NW-SE trending zone through Kennford, along an E.-W. line passing through Feniton and in the area between Exmouth and Sidmouth (fig. 4).

Convection Systems in SE Devon. Whereas negative values of the Disequilibrium Factor probably result from the downward percolation of groundwater, positive values are interpreted as indicating upflow. The spatial association between the occurrence of positive values of the Disequilibrium Factor and the distribution of high values of U and/or K suggests that, with the possible exception of a short interval during the Upper Cretaceous, groundwater was available to participate in circu-

FIG. 4. SE Devon: Soil Disequilibrium Factor Values (ψ).
 $\psi = \{ \text{Total Count (5 min)/100} \} - \{ 43.95 \times K (\%) \} - \{ 25.11 \times U (\text{ppm}) \} - \{ 11.45 \times Th (\text{ppm}) \}$.

lation systems at all times. The upflow pathways in the groundwater movement pattern for SE Devon occur along major basement fracture systems. The E.-W. zone through Feniton is coincident with the position of the faulted southern margin of the Crediton Trough, the zone between Exmouth and Sidmouth occurs along the line of a westerly extension of the Abbotsbury fault (Melville and Freshney, 1982), and the NW zone through Kennford has already been discussed in terms of the transcurrent faulting of SW England. For movement on such a large geographical scale, the presence of a widely distributed driving force is required for the maintenance of circulation. This is in keeping with the system of groundwater circulation expected to be produced by a geothermal head.

The localization of Rn anomalies along the NW-SE line through Kennford, suggests that the speed of groundwater movement up this fracture zone is more rapid than along the E.-W. lines. In turn, this may indicate some element of residual E.-W. tensional stress acting in SE Devon. Certainly the *in situ* stress anisotropy in the Carnmenellis granite is in keeping with this concept of E.-W. tension (Durrance *et al.*, 1982). However, the occurrence of the highest Rn values in the streams near the mouth of the Exe Estuary suggests that fracture flow in this area is very rapid, and some control might be exercised by formation of an intense fracture zone where NW-SE and E.-W. lines meet. Moreover, flow along a fracture zone in the pipe flow model is not two-dimensional. Along any fracture zone, areas of high upwelling activity

and areas of downflow or no activity, are to be expected. The pattern of distributions seen along the E.-W. and NW-SE lines in fig. 5 could merely reflect this variation.

FIG. 5. Summary map of SE Devon showing possible major fracture zones and areas of recent groundwater upflow.

Implications for radionuclide transport. While it is recognized that the foregoing discussion is based on investigations carried out in specific study areas in specific geological environments, it is believed that the implications of the work outlined are important in the selection of any site for the geological disposal of radioactive waste. The important conclusion in this respect is that the main control of convective groundwater circulation appears not to be a heat source producing local enhancement of the geothermal gradient, but permeability over a sufficient vertical range to allow normal geothermal gradients to act as the heat source.

Enhancement of heat flow will, however, increase the potential for convective groundwater circulation. The possibility of a thermal component to the total head in groundwater systems is therefore greater in the two belts of high heat flow in the UK outlined by Richardson and Oxburgh (1979). The first of these belts, in northern England, occupies a NW-SE trending area between Cumbria and the Wash in which a maximum heat flow of 2.51 hfu has been recorded. The second of these belts, in which heat flows as high as 3.0 hfu have been recorded, occupies an E.-W. trending area extending from the area underlain by the SW England granite batholith to the Hampshire Basin. Within these

areas permeability is still the major control of thermal groundwater movement, deeply penetrating fault-faulture systems acting as the main conduits for deep water movement.

Away from these areas of high geothermal gradient, permeability is the only control of convective water circulation. That it exists in areas of normal heat flow is exemplified by the thermal waters of the Bristol-Bath area where heat flow values are actually below average. Here deep fracturing with a long history and a strong influence of Mesozoic sedimentation in the region has allowed waters to rise from considerable depth. Residence time for water in this system has been estimated as less than 10 ka but some water could be much older (Andrews *et al.*, 1982).

Deep sedimentary basins also commonly exhibit associated geothermal phenomena. In the Paris Basin, for example, groundwater in high porosity sedimentary strata has provided the basis for a small urban heating system at Melun. In Britain, the Hampshire Basin may also have geothermal potential, as permeable sedimentary strata extend to depths of several kilometres (Garnish, 1976) while the deep sedimentary basin of Lincolnshire and E. Yorkshire is an extension of the North Sea Basin in which many high-temperature oil-field brines have been observed (Garnish, 1976). In any deep sedimentary sequence interbedding of permeable and impermeable strata presents the possibility of extensive horizontal water movement where vertical systems are interrupted by barriers of impermeable rocks.

Thus, in a number of areas in Britain, the possibility of a thermal head in addition to any topographically induced head that may be present should be considered. In an area of generally low relief (as is much of Britain) topographically induced hydraulic head values are likely to be low. Thermally induced head may therefore make a significant contribution to the total head and thus to groundwater movement. The maximum topographic head that is present in SE Devon, for example, is of the order of 250 m. If a similar calculation to that of Elder (1981) for the Wairakei geothermal field is carried out for this area, with a geothermal gradient of around 30 °C/km and a temperature differential of 150 °C maintained over a depth of 5 km, then a geothermal head in the order of 500 m results. This is a maximum value developed over that depth range. The effective head at shallower depths will be reduced by cooling effects during ascent. It should be emphasized that this head, developed on a large horizontal scale and, in the presence of deep fractures, may be very effective in producing major groundwater move-

Selection of sites for the geological disposal of radioactive waste must take into account the consequences of leakage of radionuclides from their containment. The major mechanism for the return of these toxic substances to the biosphere is transport in moving groundwater, and models are being devised to predict groundwater movement over long time scales in the region of a depository. It is normally assumed that the major driving force for water movement is topographically induced hydraulic head. This assumption may not always be justified. An additional, thermally induced head may also be present and should be taken into consideration during the early stages of site assessment.

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MODELING OF SATURATED ZONE AT YUCCA MOUNTAIN, NEVADA

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ABSTRACT

This paper presents preliminary results of groundwater modeling of the saturated zone in the vicinity of Yucca Mountain. The analysis used both a regional (approximately 250 x 250 km) and subregional (50 x 50 km) finite difference model. Numerical simulations were conducted to determine the impact of various disruptive conditions that might take place over the life span of a potential repository located beneath Yucca Mountain on the saturated groundwater flow field, as well as changes in the water-table elevations. These conditions included increases in precipitation and groundwater recharge within the regional model, changes in permeability of existing hydrogeologic barriers, and the vertical intrusion of volcanic dikes at various orientations through the saturated zone. Based on the regional analysis, the rise in the water-table in the vicinity of Yucca Mountain due to various postulated conditions ranged from only a few meters to over 200 m in certain cases. The increase in evapotranspiration and surface runoff as the water table rises was neglected in this analysis. Results of the subregional model analysis, which was used to simulate intrusive dikes approximately 4 km in length in the vicinity of Yucca Mountain, showed water-table rises ranging from a few meters to as much as 103 m. Dikes orientated approximately north-south beneath Yucca Mountain produced the highest water-table rises. Some of the results from both the regional and subregional analysis are presented in this paper. A more complete discussion of the results and details of the analysis is given in a recent report by Ahola and Sagar.¹

APPENDIX G

APPENDIX G

**Estimates of Magnitude of Rise of the Water Table Resulting from
Reduction of Hydraulic Conductivity by Earthquake Induced Strain,
Fig. G-1 through G-3**

**"Alternate Conceptual Models in the Saturated Zone at Yucca
Mountain" a presentation before the Nuclear Waste Technical
Review Board**

by L. L. Lehman and T. P. Brown

Note: For Hydrologic response in wells Ad-112 and UE-20p see Figures G-2 and G-3.

Figure G-1. Estimate of ΔH_{EG} , based on locally observed strain vs. DH relationship.

Monthly water level measurements for monitor well AD-11

Note:

- a) The monitor well Ad-11 is situated ~20 km north of Death Valley Junction, CA, and
- b) At the AD-11 site, the earthquake sequence induced strain was no larger than 10^{-8} .

Figure G-2. Maximum hydrologic response near Yucca Mountain to the Landers-Big Bear-Skull Mtn. earthquake sequence of June, 1992.

Note:

- a) The Handely event occurred on March 26, 1970;
- b) At an average distance of ~6 km, the detonation induced strain was estimated at 2×10^{-5} , Bucknam and Dickey, 1971 ;
- c) UE-20p is situated ~6 km north of the Handley G-0.

Figure G-3. Hydrologic response in well UE-20p to the Handley event. From Dudley et al., 1974.

ALTERNATE CONCEPTUAL MODELS IN THE SATURATED ZONE AT YUCCA MOUNTAIN

a presentation before the Nuclear Waste Technical Review Board
April 14, 1994

Linda L. Lehman
Tim P. Brown

Introduction

The State of Nevada has funded L. Lehman & Associates to conduct research on water level changes in the vicinity of Yucca Mountain. As a result of analyses done over the past several years, L. Lehman & Associates has made several observations. Taken together, these observations can assist in formulating a different and more complex conceptual model of the saturated zone than the models currently being analyzed in performance assessments. These observations lead to a more closely coupled hydro-tectonic concept than previously accepted.

Recently L. Lehman & Associates examined relationships between data collected on spring discharge, monitor well water levels and water levels in Devil's Hole. In studying monitor well water levels, changes were observed apparently coincident with a series of three earthquakes which took place in the southwest. On June 28, 1992 a 7.5 magnitude earthquake occurred near Landers, CA and three hours later a second 6.6 magnitude earthquake occurred near Big Bear, CA. The next day, on June 29, 1992, a 5.6 magnitude quake occurred at Little Skull Mountain, located about 23 km southeast of Yucca Mountain, near the southern boundary of the Nevada Test Site (O'Brien, 1993).

The apparent earthquake effects on well water level varied from well to well. However, the effects can be grouped into four main categories:

- well levels that exhibited an upward temporary spike
- wells that exhibited rapid upward change with an apparent long-term stabilization at a higher level
- wells that showed a downward temporary spike
- and wells that showed rapid downward change with an apparent long-term stabilization at a lower level.

Wells which show increased water levels appear to be associated with SE-NW trending fault zones while those with decreased levels are not closely related to these zones. The difference between the direction of well water level change within and outside the fault zones may indicate hydraulic relationships between these structural zones. It may also give

clues about the changes in stress that occur within hydrologic units due to earthquakes.

These observed responses suggest that some areas experience compressive strain, reducing effective pore volume and raising water levels. While other areas are experiencing tensile strain, increasing effective pore volume and lowering water levels. With the progress of site characterization at Yucca Mountain, some important questions about the basic mechanisms affecting flow in the saturated zone below Yucca Mountain have emerged. Some of these questions include:

- What is the role of faults in saturated zone flow?
- How do earthquakes affect saturated zone flow?
- How would changes in insitu stress affect water levels?
- How does geothermal heat flow affect the saturated zone flow dynamics?
- What causes the large head gradient in the water table north of Yucca Mountain?

As more information is collected, it becomes apparent that the saturated zone flow dynamics are influenced by complex interplay between the basin and range tectonics, and the local geologic structure. This presentation outlines data analyses performed at L. Lehman & Associates which suggests that there exist semi-isolated zones of groundwater flow within a larger structurally controlled system. The analyses also indicate the insufficiency of a simple conceptual model for accurately describing the saturated zone system.

A. Cosine Components

In 1990 L. Lehman & Associates examined water level data from 8 water table wells located around Yucca Mountain (Lehman, et al, 1990). The water level data were fit with a cosine function using the Fit.M program of Rice (1989). This was done to examine possible cyclic behavior in water levels and to use this information in examining relationships between the saturated zone and recharge mechanisms.

Data were analyzed in 1989. The wells referred to here are shown as Figure 1. The work involved looking at similarities in water table oscillations. It was thought that if wells were responding at similar frequencies they may be hydraulically connected. The results of this work indicated similar responses at certain wells located west of Yucca Mountain, and similar responses in wells located to the east of Yucca Mountain. What was interesting about the observation was that the similar behaving wells lined up along a NE-SW direction, a trend parallel with the major fault zones and in perpendicular with the principle tensile stress direction for the region. It also lead to the conclusion that the 2 sides of the mountain were separate hydrologic systems, only loosely connected.

It was found that water levels tended to fluctuate at 2-3 year cycles. Wells west of Yucca Mountain exhibited cycles at the long end of this range of near 1000 days, while those east had fitted periods at the short end of the range at around 880 days. This was the first indication that groundwater flow on either side of Solitario Fault may not be coupled, as is usually assumed in the existing flow models for the area.

Figure 1. Location of wells that exhibited different fitted periodicity at Yucca Mountain with circles indicating periods of 870 days and squares indicating periods near 1000 days.

Figure 2. Saturated zone water table isotherms (from Sass et al, 1988).

Figure 3. Map showing locations of features examined in Devil's Hole study.

B. INTRAVALE Work

Our analyses under the Intravale project lead us to believe that flow in the unsaturated zone could also be controlled by fractures. These analyses lead us to examine the work of Sass et al (1988), which presented temperature measurements with depth at various boreholes including subsurface water temperatures. We noticed, and presented to the Nuclear Waste Technical Review Board last summer, that the water table temperature profiles were coincident with known fracture zones under Yucca Mountain and volcanic centers. This alternative model is shown in Figure 2.

In looking at the distribution of temperature at the water table, it was apparent to Sass and others that saturated zone flow was much more complex than simple 2-dimensional uniform flow. The temperature distributions suggest localized recharge near Ghost Dance Fault and a thermal hot spot along the Solitario Fault zone. This information further complicates interpretation of the effect of local faults on groundwater flow and transport pathways. This information may be indicating that faults serve as pathways for both cool water from the surface as well as warm water upwelling from depths.

C. Devil's Hole

Most recently we examined relationships between water levels in Devil's Hole, and water levels and discharge at other springs and wells in the area. We found that Devil's Hole water levels correlated well with water levels from the AD-7 and AD-8 wells located upstream (Figure 3). But when water levels for a nearby well west of Devil's Hole, AM-7, were examined a negative correlation over time was found. Closer examination showed that most of the negative correlation was due to an opposite response to the earthquakes of June 1992, and that Devil's Hole and well AM-7 were statistically uncorrelated.

The above result was surprising, first because wells 10 miles upstream showed significant correlation with Devil's Hole while the downstream well AM-7, only a mile or so away showed no correlation. Surprising also was the fact that Devil's Hole levels showed a temporary drop, while AM-7 levels rose, in response to the earthquakes. It appeared that Devil's Hole and AM-7 were located on either side of the Stewart Valley Fault which runs through Ash Meadows, and were responding as if they existed in different hydrologic domains.

We then decided to look at all the well data from DOE's Quarterly Groundwater Data Reports from May 1992 through January 1994, which included data from 1990-1993. Specifically, we wanted to examine the direction and magnitude of water level response resulting from the June 1992 earthquakes. The data consisted of well specifications and depth to water measurements taken monthly. This means that we looked at measurements which were taken days to weeks before the quakes and measurements taken in similar time lengths afterward. At this weekly time scale, short term "seismic" effects are not seen, but rather, longer term adjustments in water level. Water level adjustments ranged from negligible to over 15 feet. Only wells with adjustments of magnitude 0.5 feet or more were plotted and their locations compared with fault traces in the area. These wells are shown

Figure 4. Locations of fault zones and monitored wells that exhibited more than 1/2 feet of sustained water level change between mid-June and mid-July 1992.

Table 1. DOE monitored wells that exhibited more than 0.5 feet of water level change between mid-June and mid-July 1992.

Wells showing decrease			Wells showing increase		
Well ID	Level change (ft)	Geologic Unit	Well ID	Level change (ft)	Geologic Unit
AD-13	-1.8	alluvium	AD-11	15.7	volcanics
CF-1	-1.4	volcanics	AD-16	5.5	alluvium
YM-1	-1.1	carbonates	AD-5	3.5	alluvium
RV-1	-0.7	carbonates	J-12	3.3	volcanics
CF-3	-0.6	alluvium	J-11	3.0	volcanics
AM-3	-0.5	alluvium	AM-7	1.3	carbonates
			AM-6	0.6	carbonates
			AD-4	0.5	alluvium

in Figure 4. Eight wells showed positive change above 0.5 feet and 6 wells had declines greater than 0.5 feet. The well responses are shown in Table 1 and an appendix at the end of this report shows water level plots for these wells.

In examining the relationship between water level changes and fault zones, we found that increases in levels correlated with transform or shear fault zones and decreased levels with extensional or normal faulting. The exact reasons for these changes are unknown but one can speculate on some potential mechanisms:

- Perhaps these responses imply that areas around the transform or shear zones may have experienced a long lasting increase in compression, thereby reducing pore volume, while extensional zones may have experienced long lasting increases in pore volume due to increased tension.
- Alternatively, these long lasting level changes might result from large scale hydraulic adjustment to locally extreme in situ stress changes. Water forced out of zones due to increased compression might "upwell" at some locations where vertical conductivity is high.
- Further, changes in temperature distributions and thermal flow rates brought on by the quakes may also play a role in changing the observed water levels.

The above analyses, rather than providing any clear-cut answers regarding the mechanisms for the observed changes, seriously questions any saturated zone model which does not consider local tectonics, especially the effects of fault zones on flow. Currently these coupled effects are not well understood.

D. Project Implications

These observations may have implications for licensing the Yucca Mountain site. First, this and other work (see Winograd and Thordarson), suggests that the saturated zone flow may be quite complex, with effects from in situ stress and earthquake activity, geothermal heat flow, as well as more obvious structural and hydraulic controls. This means that performance assessment calculations based on simple uniform 2-D flow may be inadequate.

Performance assessments should be done using a wider range of conceptual models which attempt to account for tectonic and thermal affects. They also need to incorporate extreme behavior of the system due to major tectonic events, such as large earthquakes, in order to bound travel times and mass release predictions.

There also needs to be more integration of studies relating to hydrology and tectonics. Scientists from both disciplines will need to work together to answer some of the complex questions that need resolution. For example, it appears earthquake activity will not only affect facility structural integrity, but effect the hydrologic models as well.

The apparent earthquake induced water level changes probably have caused changes in local gradients and flow directions which may increase or decrease travel time calculations. Some fault zones may also be acting as conduits for rapid flow greatly decreasing travel times based on uniform flow assumptions.

The questions raised by these observations call for collection of data on temperature distributions at the water table and with depth, especially within fracture zones. This could shed light on recharge and any upwelling phenomenon that may exist. More data on water levels over time, more densely spaced over specific areas, will also help resolve relationships between faults and flow. Pressure measurements conducted at multiple levels of boreholes would help resolve questions of vertical flow. Therefore, current test plans may have to be revised.

Overall, saturated zone flow is slowly being revealed as much more complex than originally thought. If performance assessment is to be carried out at the necessary level of confidence, it will have to keep pace with the growing level of complexity seen emerging in site characterization data.

References

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APPENDIX: Plots of water levels over time for DOE monitor wells with measurements taken immediately before and after the June 1992 earthquakes shown as black circles and other measurements shown as hollow squares.

Devil's hole daily water level measurements
including Earthquake affected data of July, 1992

Monthly water level measurements for monitor well AM-7

Monthly water level measurements for monitor well J-12

Monthly water level measurements for monitor well J-11

Monthly water level measurements for monitor well AD-11

Monthly water level measurements for monitor well AD-16

Monthly water level measurements for monitor well AM-6

Monthly water level measurements for monitor well AD-4

Monthly water level measurements for monitor well AD-5

Monthly water level measurements for monitor well AM-7

Monthly water level measurements for monitor well YM-1

Monthly water level measurements for monitor well RV-1

Monthly water level measurements for monitor well AD-13

Monthly water level measurements for monitor well CF-1

Monthly water level measurements for monitor well CF-3

Monthly water level measurements for monitor well AM-3