TECHNICAL BASIS FOR RESOLUTION OF THE IGNEOUS ACTIVITY KEY TECHNICAL ISSUE

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QUALITY OF DATA, ANALYSIS, AND CODE DEVELOPMENT

DATA: CNWRA-generated data contained within this report meet quality assurance (QA) requirements described in the CNWRA Quality Assurance Manual. Sources for other data should be consulted for determining the level of quality for those data.

ANALYSIS AND CODES: Probability models that form the basis of this report have been tested for accuracy. The calculations were checked as required by QAP-014, Documentation and Verification of Scientific and Engineering Calculations, and recorded in a scientific notebook. Probability models codes also are contained in PVHA_YM Version 1.0, which was developed under TOP-18 QA procedures. Calculations for magma flow velocities, drift fracturing, and waste-package heating models in Sections 3.2.3 and 3.2.4 are documented in CNWRA scientific notebooks.

1.0 INTRODUCTION

The Igneous Activity Key Technical Issue (IA KTI) has been defined by the U.S. Nuclear Regulatory Commission (NRC) as “predicting the consequence and probability of igneous activity affecting the repository in relationship to the overall system performance objective.” Igneous activity is the process of the formation of igneous rocks from molten or partially molten material (magma). Igneous processes are normally divided into two classes; intrusive activity, whereby magma is emplaced into preexisting rocks, and extrusive or volcanic activity, whereby magma and its associated materials rise into the crust and are deposited on the earth’s surface. The dividing line between intrusive and extrusive processes and events is at times indistinct. Dikes, which are by definition intrusive features, can break through to the earth’s surface and are responsible for many lava flows. In addition, many volcanoes first start as a dike in which flow becomes constricted to a certain location, the volcanic vent. For purposes of this report, volcanic activity is restricted to mean only those features and processes associated with the volcano and volcanic vent itself.

The main objective of work within the IA KTI is to evaluate the significance of igneous activity to repository performance by reviewing and independently confirming critical data, and evaluating and developing alternative conceptual models for estimating the probability and consequence of igneous activity at the proposed repository site. The scope of work includes reviewing various U.S. Department of Energy (DOE) documents, as well as applicable documents in the open literature, participating in meetings with DOE to discuss issues related to the KTI, observing of Quality Assurance (QA) audits of DOE, conducting independent technical investigations, and performing sensitivity studies related to igneous activity and total system performance.

The IA KTI has been factored by NRC into two subissues, which contain specific technical components. The first subissue, probability, focuses on (i) definition of igneous events, (ii) determination of recurrence rates, and (iii) examination of geologic factors that control the timing and location of igneous activity. The second subissue considers the consequences of igneous activity within the repository setting. The primary topics addressed for the second subissue are (i) definition of the physical characteristics of igneous events, (ii) determination of the eruption characteristics for modern and ancient basaltic igneous features in the Yucca Mountain Region (YMR) and analogous geologic settings, (iii) models of the effect of the geologic repository setting on igneous processes, (iv) evaluation of magma-waste package/waste form interactions, and (v) determination of volcanic deposit characteristics relevant to the consequences of igneous activity.

One of the primary objectives of the NRC prelicensing program is to direct all activities towards resolving 10 KTIs considered most important to repository performance. This approach is summarized in Chapter 1 of the staff's fiscal year (FY) 1996 Annual Progress Report (Sagar, 1997). Other chapters address each of the 10 KTIs by describing the scope of the issue and subissues, path to resolution, and progress achieved during FY1996. For the purposes of this report, “staff” shall refer to NRC and Center for Nuclear Waste Regulatory Analyses (CNWRA) staff.

Consistent with NRC regulations on prelicensing consultations and a 1992 agreement with the DOE, staff-level issue resolution can be achieved during the prelicensing consultation period; however, such resolution at the staff level would not preclude the issue being raised and considered during licensing proceedings. The three categories of issue resolution currently defined by the NRC are

- Closed—Issues are considered “closed” if the DOE approach and available information acceptably addresses staff questions such that no information beyond what is currently available will likely be required for regulatory decisionmaking at the time of initial license application.

- Closed, pending additional information—Issues are considered “closed–pending” if the NRC staff has confidence that the DOE proposed approach, together with the DOE agreement to provide the NRC with additional information, acceptably addresses NRC questions such that no information beyond that provided, or agreed to, will likely be required for regulatory decisionmaking at the time of initial license application.

- Open—Issues are considered “open” if the NRC has identified questions regarding the DOE approach or needed information, and the DOE has not yet acceptably addressed the questions or agreed to provide the necessary additional information in the license application.

An important step in the staff's approach to issue resolution is to provide DOE with feedback regarding issue resolution before license application. Acceptance criteria and review methods based on proposed 10 CFR Part 63 are developed in the Yucca Mountain Review Plan (YMRP).

This report documents the technical basis staff have used to evaluate issue resolution with the DOE. Technical basis originally contained in IA Issue Resolution Status Report (IRSR), Revision 2 (U.S. Nuclear Regulatory Commission, 1999) has been updated with new information. This information includes results of ongoing investigations at the CNWRA and DOE documents produced following the Total System Performance Assessment for the Viability Assessment (TSPA-VA). Highlighted (i.e., redlined) text in this report represents significant changes from U.S. Nuclear Regulatory Commission (1999).

2.0 RELATIONSHIP OF SUBISSUES TO DOE'S REPOSITORY SAFETY STRATEGY

The IA KTI has been defined by NRC as “predicting the consequence and probability of igneous activity affecting the repository in relationship to the overall system performance objective.” This definition is comparable but broader than the hypothesis evaluated in the DOE Repository Safety Strategy (RSS) (U.S. Department of Energy, 1998a) that “volcanic events within the controlled area will be rare and the dose consequences of volcanism will be too small to significantly affect waste isolation.” As the majority of the NRC effort has been directed toward understanding the effects of volcanic activity, the differences in the focus of the two programs have been minor. The probability and consequence subissues of the overall issue are directly incorporated in both the NRC issue and the DOE RSS. Version 3 of the DOE RSS does not include igneous activity as a principal factor for postclosure safety (CRWMS M&O, 2000a). DOE considers the development and evaluation of potential disruptive processes and events as being preliminary and insufficient to allow for identification of principal factors associated with disruptive events (CRWMS M&O, 2000a). During the August 2000 technical exchange, DOE indicated that Version 4 of the RSS would provide a basis for inclusion or exclusion of igneous events as principal factors for postclosure safety. Preliminary modeling results presented by the DOE at this technical exchange clearly indicated that igneous activity is the largest potential contributor to probability-weighted expected annual dose during the first 10,000 yr postclosure. This interpretation is supported by analyses in U.S. Nuclear Regulatory Commission (1999), which also shows that volcanic disruption is the most significant contributor to total-system risk. Although neither the DOE nor NRC analyses indicate that risks from igneous activity would exceed proposed risk-based standards, these analyses demonstrate that igneous activity significantly affects postclosure performance of the proposed repository at Yucca Mountain.

3.0 TECHNICAL BASIS FOR ISSUE RESOLUTION

This report provides a summary of the technical basis that staff have used to evaluate data and models used by the DOE to support licensing activities for Yucca Mountain. Data and models presented herein expand on the IA IRSR Revision 2 (U.S. Nuclear Regulatory Commission, 1999), and update staff interpretations of the DOE technical basis for igneous activity. Interpretations of the status of issue resolution are based on the models and data presented in this report, in addition to due consideration of information available in the literature and that provided by the DOE. Models and data in this report often represent alternatives to those proposed by the DOE. These models and data will be used by staff to independently evaluate risks from the proposed repository. In addition, alternative models and data are used to evaluate the uncertainties in DOE analyses.

3.1 PROBABILITY

DOE will need to estimate the probability of future volcanic eruptions and igneous intrusions that may affect the performance of the proposed repository. Staff will review DOE assumptions made in estimation of the probability of volcanic eruptions and igneous intrusions for consistency with known past igneous activity in the YMR and to determine if the analysis and assumptions do not underestimate effects. The following sections provide information on data and models used to evaluate the probability of igneous activity in the YMR.

3.1.1 Definition of the YMR Igneous System

3.1.1.1 Technical Basis

Acceptable probability models use past patterns of YMR igneous activity to estimate probabilities of future igneous events. Current models in the available literature for the spatial and temporal recurrence of basaltic volcanism rely on probabilistic methods (e.g., Ho, 1991; Kuntz, et al., 1986; McBirney, 1992; Wadge, et al., 1994; Connor and Hill, 1995). In these models, patterns of future activity are primarily estimated from patterns of past volcanic activity, including eruption location, frequency, volume, and chemistry. In addition, geologic processes, particularly structural deformation, have been investigated as partially controlling the distribution and timing of volcanism (Bacon, 1982; Parsons and Thompson, 1991; Connor, et al., 1992; Lutz and Gutmann, 1995; Conway, et al., 1997). Probabilistic models of volcanism at the proposed repository site should be consistent with rates and timing of past volcanism and with observations made in the YMR and other volcanic fields, regarding the relationship between igneous activity and other tectonic processes.

Basaltic igneous activity has been a characteristic of the Western Great Basin (WGB) in Nevada and California since about 12 Ma (e.g., Luedke and Smith, 1981). Although much of this activity has occurred near the boundaries of the WGB since 10 Ma (Figure 1), distributed volcanism between Death Valley, Yucca Mountain, and the Reveille Range is a well-recognized feature of the WGB (e.g., Carr, 1982). Basaltic volcanism, however, is localized in specific areas of the WGB and often shows regular spatial shifts through time (Connor and Hill, 1994). Many of the WGB basaltic volcanic fields exhibit clear spatial and temporal boundaries to igneous activity. In contrast, diffuse basaltic volcanism in the YMR is distributed over a relatively large area with often ambiguous spatial and temporal bounds (Figure 1). Defining the spatial and temporal extent of the YMR magma system is the first step in quantifying patterns of igneous activity for use in probability models. Quantitative criteria, however, do not clearly define the extent of the YMR basaltic volcanic system in space and time. For example, to date, petrogenetic relationships between <6-Ma and 6–11-Ma basalts are ambiguous, as similar composition basalts occur within each interval of time. Isotopic geochemical characteristics are distinct for ≤ 6-Ma basalts located within 40 km of the proposed repository site, which is a distance that encompasses the main YMR system. Some ≤ 6-Ma basalts within 90 km south and west of the proposed repository site, however, have the same distinct compositional characteristics and, thus, may be part of the YMR volcanic system.

Numerous attempts to define the extent of the YMR basaltic volcanic system have been based on qualitative to semi-quantitative criteria. Early workers (Vaniman, et al., 1982; Crowe, et al., 1982) concluded that basalts younger than about 9 Ma were petrologically distinct from 9- to 11-Ma basalts and, thus, constitute the igneous system of interest. Subsequent work (Crowe, et al., 1983; 1986) generally confirmed this interpretation; however, many analyzed Plio-Quaternary basalts have petrogenetic characteristics similar to some 9- to 11-Ma basalts (i.e., Crowe, et al., 1986). Crowe and Perry (1989) used similar petrogenetic arguments to define the Crater Flat Volcanic Zone (CFVZ), which is a northwest-trending zone based on the occurrence of <5-Ma volcanoes between Sleeping Butte and buried volcanoes in the Amargosa Desert (Figure 2). Smith, et al. (1990) expanded the CFVZ to include Buckboard Mesa. Numerous other subdivisions are possible, based on the pattern of <5-Ma basaltic volcanoes (e.g., Crowe, et al., 1995; Geomatrix, 1996).

In the Crater Flat basin, 5–11 Ma basalt may have been produced from the same types of igneous processes that formed the 0.08–5 Ma basalt. This relationship is important because many of the source-zone models used in Geomatrix (1996) and resulting models (e.g., CRWMS M&O, 2000b) are based on the timing and location of basalt <5 Ma. If basalt older than 5 Ma was produced by the same basic processes as the younger basalt, then the timing and location of the older basalt is relevant to defining probability models for YM. Ongoing work indicates that Miocene basalt located east of the Crater Flat basin (e.g., Jackass Flat and Little Skull Mountain) commonly contains quartz xenocrysts and disequilibrium crystallization textures such as sieved/fritted plagioclase and embayed olivine. These features represent assimilation of crustal rock, which indicates the ascending basaltic magma stagnated in the crust for a significant amount of time. These disequilibrium crystallization features are absent from Miocene basalt within the Crater Flat basin (e.g., Solitario Canyon and southern Crater Flat). In addition, trace element patterns for Miocene basalts within the Crater Flat basin are similar to patterns for younger basalts, whereas Miocene basalts located outside the Crater Flat basin have trace element patterns significantly different from younger basalt. Miocene Crater Flat basin basalts also were produced after the most significant tectonic deformation in the basin, with a stress regime that is similar to Plio-Quaternary deformation (Stamatakos, et al., 2000). Basalt 0.08–11 Ma within the Crater Flat basin appears to be derived from the same fundamental igneous processes, whereas basalt 5–11 Ma outside the basin resulted from significantly distinct igneous processes. The timing and location of all <11 Ma igneous events within the Crater Flat basin thus appear relevant to evaluating the probability of future igneous activity at YM.

Isotopic geochemical characteristics commonly are used to define the extent of basaltic igneous systems (e.g., Leeman, 1970; Farmer, et al., 1989). Isotopes of Sr and Nd are distinct for ≤ 6-Ma basalts located within 50 km of the proposed repository site (Farmer, et al., 1989; Yogodzinski and Smith, 1995; Hill, et al., 1996). In addition, Pliocene basalts in the Grapevine

Mountains, Funeral Formation, and southern Death Valley (Figure 2) also share these distinctive isotopic characteristics. These more distal basalts, however, are located in significantly different tectonic regimes than the YMR. Crustal tectonics likely influence magma ascent and eruption rates (e.g., McKenzie and Bickle, 1988). Although the distal basalts may have originated from a compositionally similar mantle, differences in tectonic history or crustal lithologies may have resulted in spatial and temporal controls on basaltic volcanism that are significantly different from the YMR. Figure 2 shows the extent of basalts that are potentially part of the YMR igneous system, based on temporal, spatial, and geochemical affinities. Although a range of geochronological techniques has been utilized in the YMR to date Quaternary basaltic features, most basalts older than about 1 Ma have been dated using standard K-Ar and $^{40}Ar/^{39}Ar$ methods (Hill, et al., 1993). These data are compiled in U.S. Nuclear Regulatory Commission (1999) and are used in subsequent probability analyses. The extent of the YMR magmatic system was also considered during the DOE-sponsored formal expert elicitation (Geomatrix, 1996). This report utilized areas that generally encompassed about the same general region as that shown on Figure 2. However, more extensive regions were often included in the background or regional recurrence rate estimates. In general, the report concluded that the <5 Ma basalts were most important to define temporal recurrence rates for the YMR. However, it appears from Geomatrix, 1996, that petrologic data and models were not used to define spatial patterns or process models. It also is not clear why the 5–11 Ma volcanics were not considered by all experts to define spatial patterns or derive process models. As a result, the areas used for the regional recurrence rate estimates do not appear to be well supported by the petrologic data and models. The significance of basaltic centers >40 km from the site to probability issues depends on the model being evaluated. Probability models that depend heavily on the timing of past events (e.g., Ho, 1992) are strongly affected by inclusion of these centers in the YMR system. Depending on the time used to calculate future recurrence rates, inclusion of the distal centers may substantially elevate or decrease the probability of future eruptions at the proposed repository site. In contrast, models that spatially define the extent of the system and evaluate the area of the system to the area of the proposed repository (e.g., Crowe, et al., 1982; Geomatrix, 1996) may exhibit a marked decrease in probability at the site due to expansion of the YMR system to accommodate distal volcanoes. Finally, the presence of the distal volcanic centers has little effect on spatio-temporal recurrence models (e.g., Connor and Hill, 1995), as distal centers are too old and too far away from the proposed repository site to strongly influence the locus of volcanism in Crater Flat basin.

3.1.1.2 Summary

Sufficient information exists on the spatial and temporal extent of the YMR basaltic system to support spatio-temporal probability models (e.g., Connor and Hill, 1995). Evaluation and acceptance of other models, however, requires assessment of the petrogenesis of 0.1–11-Ma basalt of the YMR. A reasonably conservative, working hypothesis for these assessments is that all $\leq$ 6-Ma basalt within the dashed boundaries of Figure 2 is part of the YMR igneous system. Relevant data for these volcanic centers are summarized in U.S. Nuclear Regulatory Commission (1999). In addition, some 6–11-Ma basalt within these boundaries has the same petrogenesis as $\leq$ 6-Ma basalt and, thus, may be part of the YMR igneous system of interest.

All current probability estimates for future igneous activity at the proposed repository site are based on past patterns of igneous activity in the YMR. Some parameter values or ranges used in these probability models, however, are dependent on definitions of the spatial or temporal extent of the YMR igneous system. Models that may be developed by DOE subsequent to those discussed in this report will need to be evaluated independently by NRC to assure that the parameters and definitions are internally consistent.

3.1.2 Definition of Igneous Events

3.1.2.1 Technical Basis

Although all volcanic events are associated with an intrusive event, basaltic intrusions may reach subsurface depths of less than 300 m without forming a volcano (Gudmundsson, 1984; Carter Krogh and Valentine 1995; Ratcliff, et al., 1994). Therefore, probability calculations must distinguish between volcanic (i.e., extrusive) and intrusive events in order to be applicable in repository performance and risk assessment models.

Because recurrence rates used in many probability models are sensitive to the size, duration, and area affected by igneous events, igneous event definitions must be used consistently throughout an acceptable analysis. Furthermore, differences in igneous event definitions must be considered when comparing the results of different probabilistic hazard analyses. In addition, the method used to count igneous events affects the outcome of the probability analysis. Definitions of volcanic and intrusive igneous events commonly found in the geologic literature include

- Individual, mappable eruptive units
- Episodes of vent or vent-alignment formation
- Emplacement of an igneous intrusion
- Volcanic eruption and accompanying dike injection

As discussed in the following section, igneous activity in the YMR can be categorized using each of these definitions with varying degrees of confidence.

3.1.2.1.1 Individual Eruptive Units

Definitions of volcanic events vary widely in the literature (Condit, et al., 1989; Bemis and Smith, 1993; Delaney and Gartner, 1995; Lutz and Gutmann, 1995; Connor and Hill, 1995). Ideally, volcanic events would correspond to eruptions. Unfortunately, subsequent geologic processes often obliterate evidence of previous eruptions from the geologic record (e.g., Walker, 1993). Consequently, volcanic events often have been defined as mappable eruptive units, each unit being an assemblage of volcanic products having internal stratigraphic features that indicate a cogenetic origin and eruption from a common vent (Condit and Connor, 1996). A simple definition that can be applied to young cinder cones, spatter mounds, and maars is based on morphology: an individual edifice represents an individual volcanic event (Connor and Hill, 1995). In older, eroded systems, such as Pliocene Crater Flat, evidence of vent occurrence, such as near-vent breccias or radial dikes, is required. One important advantage of this definition of volcanic events is its reliance on geological and geophysical mapping, with no requirement for geochronological data. Therefore, this definition can be applied with greater confidence than the other definitions, which require relatively precise geochronological data.

Volcanic hazard analyses using the individual vent definition for volcanic events assume all mapped volcanic units occur as independent events. The resulting probability estimate is for direct disruption of the proposed repository by a single vent-forming volcanic eruption (e.g., Connor and Hill, 1995).

Several edifices can form, however, during an essentially continuous basaltic eruptive episode. For example, three closely spaced cinder cones formed during the 1975 Tolbachik eruption (Tokarev, 1983; Magus'kin, et al., 1983). In this case, the three cinder cones represent a single eruptive event that is distributed over a larger area than represented by an individual cinder cone. The three 1975 Tolbachik cinder cones have very different morphologies and erupted adjacent to three older (Holocene) cinder cones (Braytseva, et al., 1983). Together, this group of six cinder cones forms a 5-km-long, north-trending alignment. Without observing the formation of this alignment, it likely would be difficult to resolve the number of volcanic events represented by these six cinder cones if the number of volcanic events was defined as the number of eruptions. This type of eruptive activity raises uncertainties about how a number of volcanic events represented by individual volcanoes should be assessed, even where these volcanoes are well-preserved.

Geochemical and apparent geochronological variations present at some YMR Quaternary volcanoes have been interpreted as reactivation of individual volcanoes after more than 10,000-yr quiescence (Wells, et al., 1990; Crowe, et al., 1992; Bradshaw and Smith, 1994). Results from paleomagnetic (Champion, 1991; Turrin, et al., 1991) and geochronologic (Heizler et al., 1999) studies, however, contradict this interpretation and cast doubt on the likelihood that cinder cones in the YMR have reactivated long after their original formation (Whitney and Shroba, 1991; Wells, et al., 1990, 1992; Turrin, et al., 1992; Geomatrix, 1996; Perry et al., 1998). Given the possibility of cinder cone reactivation, the number of volcanoes present in the YMR may underestimate the rate of future YMR volcanic eruptions. In the context of volcanic hazards for the proposed repository, however, the spatially dispersed character of volcanism is extremely important in calculating the probability of occurrence, whereas the reactivation of an existing cinder cone is more important in determining consequence of the activity. Thus, reactivation of cinder cones is interesting as a gauge of overall activity in the volcanic system, but, is not easily related to rates of new volcano formation.

3.1.2.1.2 Episodes of Vent or Vent-Alignment Formation

Additional investigations in other volcanic fields have demonstrated that some cinder cone alignments develop over long periods of time during multiple episodes of volcanic eruption (Connor, et al., 1992; Conway, et al., 1997), particularly where a large fault controls the locations of basaltic vents. For example, Conway, et al. (1997) found that the northern segment of the Mesa Butte fault zone in the San Francisco volcanic field, Arizona, repeatedly served as a pathway for magma ascent for at least 1 m.y. and formed a 20-km-long cinder cone alignment (Figure 3). Isotopic dates reported in Conway, et al. (1997) indicate volcanism along the northern Mesa Butte fault was episodic, and successive episodes were separated in time by as much as 400 k.y. (Figure 4). Spatial patterns of volcanism along the Mesa Butte alignment apparently were independent of field-wide trends, indicated by the large lateral shifts in volcanic loci between successive episodes (Conway, et al., 1997). These observations help clarify trends observed in the development of young, potentially active volcanic alignments. For example, the largely Holocene Craters of the Moon volcanic field, Idaho, shows similar eruption patterns characterized by multiple episodes of magmatism and frequently shifting loci of volcanism along the Great Rift (Kuntz, et al., 1986), albeit on a time scale of thousands of years. This behavior contrasts sharply with eruption patterns of other short-lived fissure eruptions, such as the Laki fissure eruption (Thordarson and Self, 1993) or the Tolbachik eruption of 1975 (Tokarev, 1983). Evidence of episodic volcanism along the Mesa Butte fault indicates independent magmatic episodes may recur along geologic structures even following periods of quiescence lasting 100 k.y. or more. Volcano alignments in the YMR, such as the Amargosa Aeromagnetic Anomaly A alignment (Connor, et al., 1997), thus, may constitute multiple volcanic events. Paleomagnetic (Champion, 1991) and radiometric dating (U.S. Nuclear Regulatory Commission, 1999) of the Quaternary Crater Flat cinder cones (Figure 5) suggests these cinder cones may have formed during a relatively brief period of time (<100,000 yr) and, therefore, may represent a single eruptive event like the Tolbachik alignment. Evidence from aeromagnetic and ground magnetic surveys (Langenheim, et al., 1993; Connor, et al., 1997) suggests that older, buried volcanoes also exist in southern Crater Flat along this alignment. Therefore, the alignment may have reactivated through time, in a manner similar to the Mesa Butte volcano alignment.

Defining aligned volcanoes of similar ages as single volcanic events effectively reduces both the total number of volcanic events in the region and the regional recurrence rate. The area affected by the entire cone alignment, however, is much greater than the areas impacted by individual cinder cones. This variation in disruption area must be propagated through the volcanic hazard analysis.

Hazard analyses defining vents and vent alignments as volcanic events are used to estimate the probability of direct disruption of the proposed repository. Primary uncertainties in those probability estimates result from uncertainty in the number and distribution of volcanic vents along alignments.

3.1.2.1.3 Emplacement of an Igneous Intrusion

Igneous events are a broader class than volcanic events in that igneous events must encompass the intrusive and extrusive components of igneous activity. The number of mapped, igneous dikes generally is not considered a reasonable definition of an igneous event because multiple dikes often are injected into the shallow crust during single episodes of igneous intrusion. Furthermore, individual dikes frequently coalesce at lower stratigraphic levels. As a result, several mapped dikes may represent a single igneous event. For example, Delaney and Gartner (1995) mapped approximately 1,700 individual dikes in the Pliocene San Rafael volcanic field, Utah (Figure 6). These dikes are associated with approximately 60 breccia zones and volcanic buds, which are interpreted as the roots of eroded, volcanic vents. Based on their mapping, Delaney and Gartner (1995) suggested that approximately 175 episodes of intrusion resulted in the emplacement of the 1,700 dikes and 60 volcanic vents, but also indicated that this grouping of mapped units was a subjective process.

In the YMR, the number of Plio-Quaternary igneous events is unknown. Based on analogy with the San Rafael volcanic field, YMR intrusive events may be a factor of two or more greater than the number of volcanic events (Delaney and Gartner, 1997). Studies in the YMR by Ratcliff, et al. (1994) and Carter Krough and Valentine (1995) have demonstrated that some Miocene basaltic igneous intrusions stagnated within several hundred meters of the surface without erupting. These basaltic dikes and sills are mapped in Miocene tuffs, similar in character and composition to those underlying Yucca Mountain. Thus, probability estimates based on the number of igneous events characterized by this approach would encompass both direct disruption of the repository with transport of waste into the accessible environment during a volcanic eruption and the indirect effects, such as canister failure during dike or sill intrusion. Additional complications arise with this definition based on the limited ability of a shallow dike to laterally transport entrained material into the volcanic conduit (e.g., Spence and Turcotte, 1985). A volcano may form outside of the repository boundary, with an associated subsurface dike that penetrates the repository directly. Although an intrusive, igneous event definition would indicate disruption of the repository, the ability of the waste to be transported laterally by the dike and dispersed into the accessible environment by the volcano would be extremely limited. The definition of an igneous event as encompassing both volcanic and intrusive components, while strictly correct from a geologic perspective, is unsuitable for application in risk assessments because of the dramatically different consequences of intrusive and extrusive igneous activity. Therefore, it is best to consider only the intrusive component of igneous events under this definition, reserving extrusive components for definitions based on vents and vent alignments.

Geomatrix (1996) combined dike emplacement and volcano formation into a single igneous event class, which had a range of annual probabilities from 10^{-10} to 10^{-7} with a mean probability of 1.5×10^{-8}. In the TSPA-VA, the DOE calculated the probability of volcanic disruption of the proposed repository site by assuming the 1.5×10^{-8} mean annual probability from Geomatrix (1996) represented the probability of a dike intersecting the repository. Using the volcanic source-zone approach in Geomatrix (1996) and assuming that 0–4 vents would form along the intersecting dike, DOE calculated the mean annual probability of volcanic disruption would thus be about 6×10^{-9} (CRWMS M&O, 1998a) (See section 3.1.7). This low probability would allow screening of volcanic disruption from scenarios considered in future DOE-TSPAs (U.S. Department of Energy, 1998b).

3.1.2.1.4 Volcanic Eruptions with Accompanying Dike Injection

An igneous event can be similarly defined in terms of the subsurface area disrupted by the intrusion of magma during a volcanic event. For example, numerous dikes in the San Rafael volcanic field were injected laterally through the shallow subsurface for hundreds of meters away from volcanic vents during volcanic eruptions (Delaney and Gartner, 1995). Uncertainties resulting from this definition of an igneous event include estimates of probable lengths and widths of dike zones associated with the formation of vents and the locations of vents along these dike zones (e.g., Hill, 1996). The effects of these laterally injected dikes on performance, however, are substantially less than the direct effects of vent formation, because of the limited ability of the waste to be directly transported to the surface along nearly the length of the dikes when compared to the transportation ability of the volcanic vent itself.

3.1.2.2 Summary

There is no one generally accepted criterion to singularly define an igneous event. Repository performance considerations, however, require that the probability of volcanic disruption is calculated discretely from the probability of intrusive disruption. All volcanic events that may penetrate the proposed repository are accompanied by a subsurface intrusion. However, intrusive events may occur without direct volcanic disruption, either because a volcano does not form at the surface or the location of the volcano is at a distance greater than the lateral transport ability of a shallow dike. Therefore, the probability of intrusive, igneous events affecting the proposed repository is at least as large as, and could be significantly larger than, the probability of volcanic disruption.

Potential intrusive and extrusive events must be considered separately because the effects on repository performance are significantly different for extrusive and intrusive processes. A volcanic, igneous event that penetrates the repository has the potential to entrain, fragment, and transport radioactive material into the subaerial accessible environment. In contrast, an intrusive, igneous event that penetrates the repository would produce thermal, mechanical, and chemical loads on engineered systems, which could impact waste-package degradation. Radioactive release associated with intrusive, igneous events is through hydrologic flow and transport, rather than through direct transport by volcanic processes. Therefore, probability calculations must distinguish between volcanic and intrusive, igneous events in order to be applicable in repository performance and risk assessment models.

3.1.3 Patterns of Igneous Activity in the YMR

3.1.3.1 Technical Basis

Previous studies of volcanism in the YMR, and elsewhere, cumulatively indicate that models describing the recurrence rate or probability of basaltic volcanism should reflect the clustered nature of basaltic volcanism and shifts in the locus of basaltic volcanism through time. Models also should be amenable to comparison with basic geological data, such as fault patterns and neotectonic stress information, that affect vent distributions on a comparatively more detailed scale. The models used to estimate future igneous activity in the YMR should either explicitly account for the following or obtain bounding estimates:

- Shifts in the locus of volcanic activity through time
- Vent clusters
- Vent alignments and correlation of vents and faults

Data from other basaltic volcanic fields may be used to test the models. Each of these spatial patterns is reviewed in this section, with emphasis on the nature of these spatial patterns in the YMR and how these compare with spatial patterns in cinder cone volcanism observed in other basaltic volcanic fields. This comparison is followed by discussions in Section 3.1.4.1 of how these spatial patterns in volcanic activity can be used to calibrate and test probabilistic volcanic hazard models for disruption of the proposed repository.

3.1.3.1.1 Shifts in the Location of Basaltic Volcanism

Spatial variation in recurrence rate of volcanism in the YMR has been suggested based on apparent shifts in the locus of basaltic volcanism from east-to-west since the cessation of caldera-forming volcanism in the Miocene Southern Nevada Volcanic Field (Crowe and Perry, 1989). Well-defined shifts in volcanism have occurred in many other basaltic volcanic fields. In the Coso volcanic field, California, Duffield, et al. (1980) found that basaltic volcanism occurred in essentially two stages. Eruption of basalts occurred over a broad area in what is now the northern and western portions of the Coso volcanic field from approximately 4 to 2.5 Ma. In the Quaternary, the locus of volcanism shifted to the southern portion of the Coso volcanic field.

Condit, et al. (1989) noted the tendency for basaltic volcanism to gradually migrate from west to east in the Springerville volcanic field between 2.5 and 0.3 Ma. Other examples of continental basaltic volcanic fields in which the location of cinder cone volcanism has migrated include the San Francisco volcanic field, Arizona, (Tanaka, et al., 1986), the Lunar Crater volcanic field, Nevada, (Foland and Bergman, 1992), the Michoacán-Guanajuato volcanic field, Mexico, (Hasenaka and Carmichael, 1985), and the Cima volcanic field, California, (Dohrenwend, et al., 1984; Turrin, et al., 1985). In some areas, such as the San Francisco and Springerville volcanic fields, migration is readily explained by plate movement (Tanaka, et al., 1986; Condit, et al., 1989; Connor, et al., 1992). In other areas, the direction of migration or shifts in the locus of volcanism does not correlate with the direction of plate movement. In either case, models developed to describe recurrence rate of volcanism or to predict the locations of future eruptions in volcanic fields need to be sensitive to these shifts in the location of volcanic activity.

Sensitivity to shifts in the locus of volcanism can be accomplished by weighing more recent (e.g., Pliocene and Quaternary) volcanic events more heavily than older (e.g., Miocene) volcanic events. Shifts in the locus of volcanism, however, also introduce uncertainty into the probabilistic hazard assessment. For example, in the Cima volcanic field, <1.2-Ma basaltic vents are located south of significantly older volcanic vents (Dohrenwend, et al., 1984; Turrin, et al., 1985). This suggests that probability models based on the distribution of older vents would not have forecast the location of subsequent (<1.2 Ma) eruptions adequately. In the Springerville volcanic field, large-scale shifts in the locus of volcanism accompanied a major geochemical change in the basalts from tholeiitic to more alkalic, suggesting that a fundamental change in petrogenesis may have affected shifts in the locus of volcanism (Condit and Connor, 1996).

As the period required for large-scale shifts in the locus of volcanism is much greater than the period of performance of a repository, the effects of these shifts can be effectively mitigated in the probability models by simply applying a greater weight to the distribution of Quaternary volcanic events than older volcanic events in the probability analysis.

3.1.3.1.2 Vent Clustering

Crowe, et al. (1992) and Sheridan (1992) noted that basaltic vents appear to cluster in the YMR. Connor and Hill (1995) performed a series of analyses of volcano distribution that yielded several useful observations about the nature of volcano clustering in the region. First, vents form statistically significant clusters in the YMR. Spatially, volcanoes younger than 5 Ma form four clusters: Sleeping Butte, Crater Flat, Amargosa Desert, and Buckboard Mesa. The Crater Flat and Amargosa Desert Clusters overlap somewhat due to the position of Lathrop Wells volcano and the three Amargosa Aeromagnetic Anomaly A vents (Figure 7). Second, a volcanic event located at the repository would be spatially part of, albeit near the edge of, the Crater Flat Cluster, rather than forming between or far from clusters in the YMR. Third, three of the four clusters reactivated in the Quaternary, indicating these clusters are long-lived and, thus, provide some constraints on the areas of future volcanism.

Cinder cones are known to cluster within many volcanic fields (Heming, 1980; Hasenaka and Carmichael, 1985; Tanaka, et al., 1986; Condit and Connor, 1996). Spatial clustering can be recognized through field observation or through the use of exploratory data analysis or cluster analysis techniques (Connor, 1990). Clusters identified using the latter approach in the Michoacán-Guanajuato and the Springerville volcanic fields were found to consist of 10–100 individual cinder cones. Clusters in these fields are roughly circular to elongate in shape with diameters of 10 to 50 km. The simplest explanation for the occurrence, size, and geochemical differences between many of these clusters is that these areas have higher magma supply rates from the mantle. Factors affecting magma pathways through the upper crust, such as fault distribution, appear to have little influence on cluster formation (Connor, 1990; Condit and Connor, 1996). In some volcanic fields, such as Coso, the presence of silicic magma bodies in the crust may influence cinder cone distribution by impeding the rise of denser mafic magma (Eichelberger and Gooley, 1977; Bacon, 1982), resulting in the formation of mafic volcano clusters peripheral to the silicic magma bodies.

Basaltic vent clustering has a profound effect on estimates of recurrence rate of basaltic volcanism. For example, Condit and Connor (1996) found that recurrence rate varies by more than two orders of magnitude across the Springerville volcanic field due to spatio-temporal clustering of volcanic eruptions. In the YMR, Connor and Hill (1995) identified variations in recurrence rate of more than one order of magnitude from the Amargosa Desert to southern Crater Flat due to the clustering Quaternary volcanism. In contrast, probability models based on a homogeneous Poisson density distribution that ignores clustering will overestimate the likelihood of future igneous activity in parts of the YMR far from Quaternary centers and underestimate the likelihood of future igneous activity within and close to Quaternary volcano clusters.

3.1.3.1.3 Vent Alignments and Correlation of Vent Alignments and Faults

Tectonic setting, strain-rate, and fault distribution all may influence the distribution of basaltic vents within clusters, and sometimes across whole volcanic fields (Nakamura, 1977; Smith, et al., 1990; Parsons and Thompson, 1991; Takada, 1994). Kear (1964) discussed local vent alignments, in which vents are the same age and easily explained by a single episode of dike injection, and regional alignments, in which vents of varying age and composition are aligned for distances 20–50 km or more. For example, by Kear's (1964) definition, the Mesa Butte alignment (Figure 3) would be a regional alignment that is more likely to reactivate after a long period of quiescence than a local alignment. Thus, this distinction between local and regional alignments can potentially alter probability estimates.

Numerous mathematical techniques have been developed to identify and map vent alignments on different scales, including the Hough transform (Wadge and Cross, 1988), two-point azimuth analysis (Lutz, 1986), frequency-domain map filtering techniques (Connor, 1990), and application of kernel functions (Lutz and Gutmann, 1995). Regional alignments identified using these techniques are commonly colinear or parallel to mapped regional structures. For example, Draper, et al. (1994) and Conway, et al. (1997) mapped vent alignments in the San Francisco volcanic field that are parallel to, or colinear with, segments of major fault systems in the area. About 30 percent of the cinder cones and maars in the San Francisco volcanic field are located along these regional alignments (Draper, et al., 1994). Lutz and Gutmann (1995) identified similar patterns in the Pinacate volcanic field, Mexico. Although alignments clearly can form as a result of single episodes of dike injection (Nakamura, 1977) and, therefore, are sensitive to stress orientation (Zoback, 1989), there are also examples of injection along pre-existing faults (e.g., Kear, 1964; Draper, et al., 1994; Conway, et al., 1997). Therefore, stress orientation in the crust and orientations of faults are indicators of possible vent-alignment orientations.

In the YMR, Smith, et al. (1990) and Ho (1992) define north-northeast-trending zones within which average recurrence rates exceed that of the surrounding region. The trend of these zones corresponds to cinder cone alignment orientations, including Quaternary Crater Flat and Sleeping Butte, that Smith, et al. (1990) and Ho (1992) hypothesize may occur as a result of structural control. Recent geophysical surveys of Amargosa Aeromagnetic Anomaly A provide further evidence of the significance of northeast-trending alignments in the YMR (Connor, et al., 1997). The ground magnetic map of data collected over Amargosa Aeromagnetic Anomaly A delineates three separate anomalies associated with shallowly buried basalt with a strong reversed polarity remnant magnetization (Figure 7). These anomalies are distributed over 4.5 km on a northeast trend, each having an amplitude of 70–150 nT. Although these features can be partially resolved with aeromagnetic data (Langenheim, et al., 1993), trenchant details emerge from the ground magnetic survey that are important to probabilistic volcanic hazard analyses and tectonic studies of the region. The southernmost anomaly, which has a smaller amplitude than those to the north but is nonetheless distinctive, and the northeast-trending structure within the negative portion of the central anomaly, which mimics the overall trend of the alignment (Figure 7), are important characteristics. The ground magnetic data also enhance the small positive anomalies north of each of the three larger-amplitude, negative anomalies, reinforcing the interpretation that Amargosa Aeromagnetic Anomaly A is produced by coherent basaltic vents with strongly reversed remnant magnetization.

A key result of this ground magnetic survey is identification of the northeast trend of the anomalies, which is quite similar to the alignment of five Quaternary cinder cones in Crater Flat (Figure 5) and to the Sleeping Butte cinder cones, a Quaternary vent alignment 40 km to the northwest of Crater Flat. Although the age of the Amargosa Aeromagnetic Anomaly A alignment is at present uncertain, it suggests that development of northeast-trending cone alignments is a pattern of volcanism that has persisted through time in the YMR and supports the idea that future volcanism may exhibit a similar pattern (Smith, et al., 1990).

Other ground magnetic surveys provide further evidence of cinder cone localization along faults (Stamatakos, et al., 1997a; Connor, et al., 1997). Northern Cone is located approximately 8 km from the repository site in Crater Flat and is the closest Quaternary volcano to Yucca Mountain. Its proximity to the site of the proposed repository makes the structural setting of Northern Cone of particular interest to volcanic hazard assessment. Northern Cone consists of approximately 0.4 km^2 of highly magnetized (10–20 A m^{-1}) lava flows, near-vent agglutinate, and scoria aprons resting on a thin alluvial fan. Large-amplitude, short-wavelength magnetic anomalies were observed over the lavas. No evidence of northeast-trending structures was discovered that could directly relate Northern Cone to the rest of the Quaternary Crater Flat cinder cone alignment. Instead, prominent linear anomalies surrounding Northern Cone trend nearly north-south and have amplitudes of up to 400 nT (Figure 8). These anomalies likely result from offsets in underlying tuff across faults extending beneath the alluvium (cf. Faulds, et al., 1994).

The relationship between faults and Northern Cone is clarified when the ground magnetic map is compared with topographic and fault maps (Frizzell and Schulters, 1990; Faulds, et al., 1994). The north-trending anomalies at Northern Cone roughly coincide with mapped faults immediately north of the survey area that have topographic expression resulting from large vertical displacements. These mapped faults and faults inferred from the magnetic map are all oriented north to north-northeast, which are trends favorable for dilation and dike injection in the current stress state of the crust (e.g., Morris, et al., 1996). Thus, the Northern Cone magnetic survey provides further support for the concept that volcanism on the eastern margin of Crater Flat was localized along faults.

Thus, there is ample evidence to suggest patterns in YMR basaltic volcanic activity are influenced by the stress state of the crust and by fault patterns. This influence includes the development of northeast-trending volcanic alignments and the localization of vents along faults. Smith, et al. (1990) noted that the occurrence of northeast-trending alignments is particularly important because much of the Quaternary volcanic activity in the region has occurred southwest of the proposed repository site. Furthermore, faults that bound and penetrate the repository block have a map pattern similar to those faults that have hosted volcanism at Northern Cone and Lathrop Wells. Given these observations, probability models for igneous disruption of the proposed repository need to account for these trends because they tend to increase the probability of igneous activity at the site relative to spatially homogenous models.

3.1.3.2 Summary

Good agreement exists on the basic patterns of basaltic volcanism in the YMR. These patterns include changes in the locus of volcanism with time, recurring volcanic activity within vent clusters, formation of vent alignments, and structural controls on the locations of cinder cones. Each of these patterns in vent distribution has an important impact on volcanic probability models and is considered in current NRC, DOE, and State of Nevada probability models.

3.1.4 Probability Model Parameters

3.1.4.1 Technical Basis

Models to estimate the probability of volcanic disruption of the proposed repository are likely to rely on a set of parameters. Use of values or ranges for these parameters must be justified using geologic data and analyses. In the following, current understanding of parameters related to

- Temporal recurrence rate of volcanism
- Spatial recurrence rate of volcanism
- Area affected by volcanic and igneous events are discussed and evaluated

3.1.4.1.1 Temporal Recurrence Rate

Probability models use estimates of the expected regional recurrence rate of volcanism in the YMR in order to calculate the probability of future disruptive volcanic activity. Previous estimates have relied on past recurrence rates of volcanism as a guide to future rates of volcanic activity. This approach has yielded estimates of regional recurrence rate between 1 and 12 volcanic events per million years (v/m.y.) (e.g., Ho, 1991; Ho, et al., 1991; Crowe, et al., 1992; Margulies, et al., 1992; Connor and Hill, 1995), with the various definitions of what constitutes a volcanic event accounting for at least part of this range.

The simplest approach to estimate regional recurrence rate is to average the number of volcanic events that have occurred during some time period of arbitrary length. For instance, Ho, et al. (1991) average the number of volcanoes that have formed during the Quaternary (1.6 m.y.) to calculate recurrence rate. Through this approach, they estimate an expected recurrence rate of 5 v/m.y. Crowe, et al. (1982) averaged the number of new volcanoes over a 1.8-m.y. period. Crowe, et al. (1992) considered the two Little Cones to represent a single volcanic event, and, therefore, concluded that there are seven Quaternary volcanic events in the YMR. This lowers the estimated recurrence rate to approximately 4 v/m.y. The probability of a new volcano forming in the YMR during the next 10,000 yr is 4–5 percent, assuming a recurrence rate of between 4 and 5 v/m.y.

An alternative approach is the repose time method (Ho, et al., 1991). In this method, a recurrence rate is defined using a maximum likelihood estimator (Hogg and Tanis, 1988) that averages events during a specific period of volcanic activity:

$$\lambda_t = \frac{(N-1)}{(T_o - T_y)} \qquad (1)$$

where N is the number of events, T_o is the age of the first event, T_y is the age of the most recent event, and λ_t is the estimated recurrence rate. Using eight Quaternary volcanoes as the number of events, N, and 0.1 Ma for the formation of Lathrop Wells (U.S. Nuclear Regulatory Commission, 1999), the estimated recurrence rate depends on the age of the first Quaternary volcanic eruption in Crater Flat. Using a mean age of 1.0 Ma (Appendix A) yields an expected recurrence rate of approximately 8 v/m.y. The ages of Crater Flat volcanoes, however, are currently estimated at approximately 1.0 ± 0.2 Ma (U.S. Nuclear Regulatory Commission, 1999). Within the limits of this uncertainty, the expected recurrence rate is between approximately 7 and 10 v/m.y. Of course, using different definitions of volcanic events leads to different estimates of recurrence rate. For example, using the formation of vents and vent alignments during the Quaternary, $N = 3$ and the recurrence rate is 2–3 v/m.y. The repose-time method has distinct advantages over techniques that average over an arbitrary period of time because it restricts the analysis to a time period that is meaningful in terms of volcanic activity. In this sense, it is similar to methods applied previously to estimate time-dependent relationships in active volcanic fields (Kuntz, et al., 1986). Application of these methods has shown that steady-state recurrence rates characterize many basaltic, volcanic fields.

Ho (1991) applied a Weibull-Poisson technique (Crow, 1982) to estimate the recurrence rate of new volcano formation in the YMR as a function of time. Ho (1991) estimates $\lambda(t)$ as

$$\lambda(t) = \left(\frac{\beta}{\theta}\right)\left(\frac{t}{\theta}\right)^{\beta-1} \qquad (2)$$

where t is the total time interval under consideration (such as the Quaternary), and β and θ are intensity parameters in the Weibull distribution that depend on the frequency of new volcano formation within the time period, t. In a time-truncated series, β and θ are estimated from the distribution of past events. In this case, there are $N = 8$ new volcanoes formed in the YMR during the Quaternary. β and θ are given by (Ho, 1991):

$$\beta = \frac{N}{\sum_{i=1}^{N} ln\left(\frac{t}{t_i}\right)} \quad (3)$$

and

$$\theta = \frac{t}{N^{1/\beta}} \quad (4)$$

where t_i refers to the time of the i^{th} volcanic event. If β is approximately equal to unity, there is little or no change in the recurrence rate as a function of time, and a stationary nonhomogeneous Poisson model would provide an estimate of regional recurrence rate quite similar to the nonhomogeneous Weibull-Poisson model. If β>1, then a temporal trend exists in the recurrence rate and the system is waxing; new volcanoes form more frequently with time. If β<1, new volcanoes form less frequently over time, and the magmatic system may be waning.

Where few data are available, such as in analysis of volcanism in the YMR, the value of β can be strongly dependent on the period *t* and the timing of individual eruptions. This dependence strongly reduces the confidence with which β can be determined. Ho (1991) analyzed volcanism from 6 Ma, 3.7 Ma, and 1.6 Ma to the present and concluded that volcanism is developing in the YMR on time scales of *t* = 6 Ma and 3.7 Ma, and has been relatively steady, β = 1.1, during the Quaternary.

Uncertainty in the ages of Quaternary volcanoes has a strong impact on recurrence rate estimates calculated using a Weibull-Poisson model. For example, if mean ages of Quaternary volcanoes are used (U.S. Nuclear Regulatory Commission, 1999) and *t* = 1.6 Ma then, using Ho (1991), β = 1.1. The probability of a new volcano forming in the region within the next 10,000 yr is thus approximately 5 percent. This value agrees well with recurrence rate calculations based on simple averaging of the number of new volcanoes that have formed since 1.6 Ma.

Crowe, et al. (1995), however, concluded that the Weibull-Poisson model is strongly dependent on the value of *t* and suggested that *t* should be limited to the time since the initiation of a particular episode of volcanic activity. This has an important effect on Weibull-Poisson probability models. If mean ages of Quaternary volcanoes are used and *t* = 1.2 Ma, the probability of a new volcano forming in the next 10,000 yr drops from 5 percent to 2 percent, and β<1, indicating waning activity. Alternatively, if volcanism was initiated along the alignment approximately 1.2 Ma but continued through 0.8 Ma, the expected recurrence rate is again close to 5 v/m.y., and the probability of new volcanism in the YMR within the next 10,000 yr is about 5 percent (*t* = 1.2 Ma). The confidence intervals calculated on $\lambda(t)$ are quite large in all of these examples due to the few volcanic events (*N* = 8) on which the calculations are based (Connor and Hill, 1993).

Cumulatively, these analyses indicate that a broad range of recurrence rates should be considered, this range varying with the definition of igneous event used. Many recurrence rate models depend on additional information to estimate recurrence rates of volcanism. Bacon (1982) observed that cumulative-erupted volume in the Coso volcanic field since about 0.4 Ma is remarkably linear in time. Successive eruptions occur at time intervals that depend on the cumulative volume of the previous eruptions. This linear relationship was used by Bacon (1982) to forecast future eruptions and to speculate about processes, such as strain rate, that may govern magma supply and output in the Coso volcanic field. Kuntz, et al. (1986) successfully applied a volume-predictable model to several areas on the Snake River Plain, where recurrence rates of late Quaternary volcanism are much higher than in the Coso volcanic field, but the cumulative volumetric rate of basaltic magmatism is, nonetheless, linear in time. Condit and Connor (1996) discovered volume eruption rates were relatively constant in the Springerville volcanic field between 1.2 and 0.3 Ma, but the number of cone-forming eruptions varied in time, in conjunction with changes in petrogenesis. These relationships between eruption volume, petrogenesis, strain rate, and frequency of volcanic events observed in other volcanic fields suggest that recurrence rate estimates in the YMR can be further refined by considering fault location, magma generation, and strain rate.

A recent paper by Wernicke, et al. (1998) has suggested that the strain rates in the Yucca Mountain area are at least an order of magnitude higher than would be predicted from the Quaternary volcanic and tectonic history of the area. Wernicke, et al. (1998) further suggest that because of what they consider anomalous strain in the Yucca Mountain area, the current probabilities of future magmatic and tectonic events may be underestimated by an order of magnitude. Based on analysis of available information, staff conclude that several alternative interpretations are possible for the strain-rate data presented by Wernicke, et al. (1998). These alternative interpretations do not result in an increase in volcanic recurrence rate (Connor, et al., 1998). It is the NRC's understanding that DOE will be funding studies to determine if the strain rates observed by Wernicke, et al. (1998) can be verified.

Subsequent to the release of the paper by Wernicke, et al. (1998) NRC received a copy of a study by Earthfield Technology, Inc., (Earthfield Technology, 1995) from DOE that provides processing and interpretation of the available regional gravity and aeromagnetic data. Appendix II of Earthfield Technology (1995) contains a map that shows the locations of 42 aeromagnetic anomalies that are interpreted as buried intrusions in the Yucca Mountain area (Figure 9). These anomalies cannot be correlated with previously recognized volcanic centers buried beneath alluvium in the Amargosa Desert (Langenheim, et al., 1993; Connor, et al., 1997). As part of ongoing uncertainty analyses, CNWRA staff conducted 12 ground magnetic surveys (Figure 9) over aeromagnetic anomalies with characteristics suggesting buried basalt (Magsino, et al., 1998). Two of these surveys encountered features consistent with small, buried basaltic centers, coincident with Earthfield Technology (1995) interpretations (features E1 and E2, Figure 9). A third survey, coincident with an Earthfield Technology (1995) anomaly (feature E3), imaged faulted tuffaceous bedrock.

Earthfield Technology (1995) interpreted 6 buried intrusions within about 5 km of the proposed repository site. If these anomalies represented basaltic igneous features, their relative proximity to the proposed repository site could affect probability models significantly. Although these anomalies have not been investigated with ground magnetic surveys, CNWRA surveys east of the proposed repository site (Figure 9) mapped features consistent with faulted tuffaceous bedrock (Magsino, et al., 1998). The proximity of these six Earthfield Technology (1995) anomalies to surface exposures of tuff, their limited extent, and overall magnetic characteristics are very similar to anomalies east of the repository site investigated by Magsino, et al. (1998). Although these six Earthfield Technology (1995) anomalies are most likely caused by faulted tuffaceous bedrock, the limited available data cannot preclude some relationship to buried igneous features.

Earthfield Technology (1995) anomalies east of 560000E and near the Funeral Mountain foothills (Figure 9) may be related to nearby surface exposures of Miocene basalt. These possible buried basaltic features, however, are too old and too distant from the proposed repository site to affect probability models significantly. Several Earthfield Technology (1995) anomalies in southern Crater Flat may possibly relate to nearby surface exposures of 11.2 Ma basalt. Based on comparison with anomalies surveyed by Magsino, et al. (1998), these and other nearby Earthfield Technology (1995) anomalies are most likely caused by faulted tuffaceous bedrock.

Although NRC has no independent basis for disagreeing with the strain-rate data presented by Wernicke, et al. (1998), it recognizes that other interpretations of these data can be made that do not require any change in the volcanic hazard assessment for Yucca Mountain (Connor, et al., 1998). The results of Earthfield Technology (1995) and Magsino, et al. (1998), however, could be used to support the arguments of Wernicke, et al. (1998) that volcanic recurrence rates are greater than currently estimated.

CRWMS M&O (2000b) concluded that the data in Earthfield Technology (1995) were mislocated during the original surveys and thus no interpretations of buried igneous features are possible from these data. Anomalies within 5 km of the proposed repository site would affect many probability models used in Geomatrix (1996), if these anomalies represented buried basaltic igneous features. DOE will need to demonstrate that recurrence rates used in licensing accurately reflect the number and timing of past igneous events in the YMR. New aeromagnetic surveys were conducted recently over the YMR (Blakley, et al., 2000). DOE has committed to evaluate these data for potential buried igneous features, if the resolution of the data is sufficient to warrant such an evaluation. This evaluation is necessary to provide reasonable assurance that all appropriate igneous features have been used to determine recurrence rate parameters.

3.1.4.1.2 Spatial Recurrence Rate

Early models assessing the probability of future volcanism in the YMR and the likelihood of a repository-disrupting igneous event relied on the assumption that Plio-Quaternary basaltic volcanoes are distributed in a spatially uniform, random manner over some bounded area (e.g., Crowe, et al., 1982; Crowe, et al., 1992; Ho, et al., 1991; Margulies, et al., 1992). However, as discussed in Section 3.1.4, patterns in the distribution and age of basaltic volcanoes in the YMR make the choice of these bounded areas subjective. For example, Smith, et al. (1990) and Ho (1992) define north-northeast-trending zones within which average recurrence rates exceed that of the surrounding region. These zones correspond to cinder cone alignment orientations that Smith, et al. (1990) and Ho (1992) hypothesize may result from structural control. These narrow zones lead to comparatively high estimates of spatial recurrence rate and probability of volcanic disruption of the proposed repository site. Utilizing bounded areas that are large compared to the current distributions of cinder cone clusters, however, results in relatively low estimates of spatial recurrence rate. Ho (1992) argued that, under these circumstances, using narrow bounding areas that include the proposed repository gives conservative estimates of probability of volcanic disruption.

Alternatively, spatial recurrence rate can be estimated using models that explicitly account for volcano clustering (Connor and Hill, 1995). This approach features several characteristics of nearest-neighbor methods that make them amenable to volcano distribution studies and hazard analysis in areal volcanic fields. First, volcanic eruptions, such as the formation of a new cinder cone, are discrete in time and space. Using nearest-neighbor methods, the probability surface is estimated directly from the location and timing of these past, discrete volcanic events. As a result, nearest-neighbor models are sensitive to patterns generally recognized in cinder cone distributions. Resulting probability surfaces also are continuous, rather than consisting of abrupt changes in probability that must be introduced in spatially homogeneous models. Continuous probability surfaces can be readily compared to other geologic data, such as fault locations, that may influence volcano distribution. Nearest-neighbor methods also eliminate the need to define areas or zones of volcanic activity, as is required by all spatially homogeneous Poisson models.

Past volcanic activity can be used to estimate parameters used in these spatially nonhomogeneous Poisson probability models for disruption of the proposed repository. This is particularly important in modeling the distribution of volcanism in the YMR because of vent clustering. As discussed previously (Section 3.1.2.3), vent clustering results in dramatic changes in spatial recurrence rate across the YMR. In order to model clustering and use these models in the probabilistic volcanic hazards assessment (PVHA), it is necessary to estimate parameters used in the models. One approach to parameter estimation is to use observed volcano distributions as a basis for comparison. This parameter estimation can be done formally, if appropriate models are used.

One estimation method for the spatial recurrence rate of volcanic events in the YMR and the probability of future volcanic events uses kernel or weighting functions (Silverman, 1986; Lutz and Gutmann, 1995; Connor and Hill, 1995; Condit and Connor, 1996). In volcanic hazard analysis, the kernel function must be estimated and used to deduce a probability density function for spatial recurrence rate of volcanism. Several types of kernels, including Gaussian and Epanechnikov kernels, are discussed by Silverman (1986). All multivariate kernels have the property

$$\int_{\mathbb{R}} K(\mathrm{x})\ d\mathrm{x} = 1 \qquad (5)$$

where $K(\mathbf{x})$ is the kernel function, and **x** is an *n*-dimensional vector in real space $\mathbb{R}$. A Gaussian kernel function for 2D spatial data is

$$K(x,y) = \frac{1}{2\pi} \exp\left\{ -\frac{1}{2}\left[(x - x_v)^2 + (y - y_v)^2 \right] \right\} \qquad (6)$$

where the kernel is calculated for a point x, y and the center of the kernel, in this case the volcano location, is x_v, y_v. The kernel is normalized using the smoothing parameter, h, making the kernel a Gaussian function, where h is equivalent to the standard deviation of the distribution:

$$K(x,y) = \frac{1}{2\pi h^2} \exp\left\{ -\frac{1}{2}\left[\left(\frac{x - x_v}{h}\right)^2 + \left(\frac{y - y_v}{h}\right)^2 \right] \right\} \tag{7}$$

If x and y locations are on a rectangular grid, the probability density function based on the distribution of N volcanoes is

$$\hat{f}(x,y) = \frac{1}{N} \sum_{i=1}^{N} K(x,y) \tag{8}$$

The above equations can be used to estimate spatial recurrence rate of volcanism, or the probability of volcanic disruption of the proposed repository site, given a volcanic eruption in the region. The results of this probability estimate depend on h. The approach to bounding uncertainty in the probability estimates resulting from this calculation is to evaluate probability using a wide range of h (Connor and Hill, 1995). Alternatively, the effectiveness of the kernel model and optimal values of h can be deduced from the distribution of nearest-neighbor distances between existing volcanoes. For example, the 2D-Gaussian kernel model can be compared with the distribution of nearest-neighbor distances between existing volcanoes by recasting the kernel function (Eq. 7) in polar coordinates:

$$K(r,\theta) = \frac{2}{h(2\pi)^{3/2}} \exp\left[-\frac{1}{2}\left(\frac{r^2}{h^2} \right) \right] \tag{9}$$

where r, θ is distance and direction from the nearest-neighbor volcanic event. The cumulative probability density function then becomes

$$\hat{F}(R) = \int_0^{2\pi}\int_0^{R} \frac{2}{h(2\pi)^{3/2}} \exp\left[-\frac{1}{2}\left(\frac{r^2}{h^2} \right) \right] dr d\theta \tag{10}$$

where $\hat{F}(R)$ is the expected fraction of volcanic events within a distance R of their nearest-neighbor volcanic event.

Distance to nearest-neighbor volcanic event in the YMR varies, depending on the definition used for a volcanic event. Treating all vents as individual volcanic events, the mean distance to nearest-neighbor volcanic event is 3.8 km with a standard deviation of 5.8 km. Some vents, such as southwest and northeast Little Cones, however, are quite closely spaced and may be treated as single volcanic events. Treating vents spaced more closely than 1 km as single volcanic events, the mean distance to nearest-neighbor volcanic event increases to 5.0 km and the standard deviation to 5.9 km. Alternatively, volcanic events can be defined in terms of vents and vent alignments. In this definition, Quaternary Crater Flat volcanoes are taken as a single event, as is Pliocene Crater Flat. Using this definition, mean distance to nearest-neighbor volcanic event increases to 7.0 km with a standard deviation of 6.4 km.

The observed fraction of volcanoes erupted at a given nearest-neighbor distance or less is compared with a Gaussian kernel model with standard deviations of 3–7 km in Figure 10. A Gaussian kernel model with h = 5 km reasonably describes the expected distance to nearest-neighbor volcano, particularly between 5 and 10 km. Smaller values, such as h = 3 km, model the distribution of individual vents at distances less than 4 km, but do not compare well with vent distributions at distances greater than 4 km. For instance, the h = 3 km model predicts that 95 percent of all volcanoes will be located at nearest-neighbor distances less than 6 km, but actually 15–40 percent of all volcanoes in the YMR are located at greater distances than this, depending on the definition of volcanic events used. The h = 7 km model tends to slightly overestimate the number of volcanoes at nearest-neighbor distances greater than 8 km. Thus, the h = 5 km model best describes the overall distribution of YMR vents and vent pairs for use in evaluation of hazards at the repository, located approximately 8 km from the nearest Quaternary volcano. This is slightly less than the standard deviation of the observed distribution, because Buckboard Mesa, located 25 km from its nearest-neighbor, is an outlier in the observed volcano distribution and increases the variance.

Vents and vent alignments have fewer nearest-neighbors than expected at distances less than 4 km if this distribution is modeled using a Gaussian kernel (Figure 10). Rather, this distribution can be modeled using a simple modification of the Gaussian kernel to account for a mean offset of the probability density function from zero:

$$\hat{F}(R) = \int_0^{2\pi}\int_0^{R} \frac{2}{h(2\pi)^{\frac{3}{2}}} \exp\left[-\frac{1}{2}\left(\frac{(r-\overline{x})^2}{h^2}\right)\right] dr d\theta \qquad (11)$$

where $\overline{x}$ is the mean offset. Incorporating a mean offset of 5–7 km and h = 3 km results in an improved fit between the observed distribution of distance to nearest-neighbor volcanic events and the Gaussian kernel model (Figure 11). The need for this mean offset arises because vent alignments are more widely spaced than individual vents. Variance does not increase significantly as a result of this increased spacing, however, when vent alignments are considered as single volcanic events. This comparatively low variance suggests there is a characteristic nearest-neighbor distance of 5–10 km in the YMR for volcanic events defined as vents or vent alignments.

This analysis indicates volcanic event distribution can be modeled using a Gaussian kernel with $h \geq 5$ km provided volcanic events are defined as individual vents or vent pairs. When vent alignments are considered as individual volcanic events, the value of h must increase to $h \geq 7$ km or the Gaussian kernel needs to be modified to include an offset distance. Thus, model testing indicates that the types of kernels and parameters used within each kernel to evaluate probability should vary with the definition of volcanic event. The Epanechnikov kernel function is widely used to estimate spatial recurrence rate in basaltic volcanic fields (Lutz and Gutmann, 1995; Connor and Hill, 1995; Condit and Connor, 1996) and may be tested in a similar manner as the Gaussian kernel function. The Epanechnikov kernel in 2D-Cartesian coordinates is

$$K_e(x,y) = \frac{2}{\pi h^2}\left\{1 - \left[\left(\frac{x - x_v}{h}\right)^2 + \left(\frac{y - y_v}{h}\right)^2\right]\right\} \qquad (12)$$

where

$$\sqrt{(x-x_v)^2 + (y-y_v)^2} \leq h$$

otherwise,

$$K_e(x,y) = 0$$

In polar coordinates this kernel function becomes

$$K_e(r,\theta) = \frac{3}{4\pi h}\left[1 - \left(\frac{r^2}{h^2}\right)\right], \quad r \leq h \qquad (13)$$

where r is distance from the volcano and θ is direction. The cumulative probability density function is then

$$\hat{F}(R) = \int_0^{2\pi}\int_0^{R}\frac{3}{4\pi h}\left[1 - \left(\frac{r^2}{h^2}\right)\right]dr d\theta, \; R \leq h \qquad (14)$$

As was accomplished with the Gaussian kernel, the cumulative probability density function for the Epanechnikov kernel can be compared with the observed fraction of volcanoes erupted at a given nearest-neighbor distance or less for various values of h (Figure 12). This comparison indicates an Epanechnikov kernel function with h = 10 km best models the distribution of distance to nearest-neighbor volcanic events, if volcanic events are defined as vents or vent pairs. If volcanic events are defined as vents or vent alignments, 15 km $<h<$18 km better approximates the distribution of distances to nearest-neighbor volcanic events, given the distribution of YMR volcanoes. Comparison of the Epanechnikov and Gaussian kernel models suggests the Gaussian kernel models better fit the observed volcano distribution than Epanechnikov distributions, particularly at nearest-neighbor distances greater than 6 km. The difficulty fitting the observed distributions with the Epanechnikov kernel function results from truncation of this distribution at distances greater than h.

Testing models against observed distributions also leads to a natural definition of conservatism. For example, the distance between the proposed repository and its nearest-neighbor Quaternary volcano is 8.2 km. A Gaussian kernel function with $h \geq$ 7 km clearly is conservative because a greater fraction of volcanic events occur at nearest-neighbor distances less than

8.2 km than predicted by the model, whereas a Gaussian kernel function with h = 3 km is not conservative (Figure 10). Similarly, probability models based on Epanechnikov kernel functions and $h \geq$ 10 km are conservative where volcanic events are defined as vents and vent pairs, and $h \geq$ 18 km where volcanic events are defined as vents and vent alignments.

3.1.4.1.3 Area Affected by Igneous Events

The area affected by igneous events varies with the definition of igneous event (Section 3.1.2). Where igneous events are defined in terms of individual, mappable eruptive units, the resulting probability estimate is for direct disruption of the proposed repository and release of waste into the accessible environment. The probability of a volcanic event disrupting the repository depends on the repository area potentially disrupted by flow of magma through the subsurface conduit of the volcano as the eruption develops. Observations at cinder cones in the process of formation (e.g., Luhr and Simkin, 1993; Fedotov, 1983; Doubik, et al., 1995) are that these eruptions initiate from dike injection at comparatively low ascent velocities, on the order of 1 m s^{-1}, which can deform an area of the ground surface several hundred meters in length. Basaltic eruptions, however, quickly localize into vent areas as the eruption progresses and magma flow velocities increase to around 100 m s^{-1}. Hill (1996) reviewed literature on subsurface areas disrupted by basaltic volcanoes analogous to past volcanic eruptions in the YMR. Based on this review and data collected at Tolbachik volcano, Russia, Hill (1996) concluded that typical subsurface conduit diameters are between 1 m and 50 m at likely repository depths of about 300 m. Vent conduits exposed in the San Rafael volcanic field (Delaney and Gartner, 1995), however, often have diameters on the order of 100 m. Therefore, areas disrupted by vent formation, potentially leading to the release of waste into the accessible environment, are on the order of 0.01 km^2 or less. Conservatively, such a volcanic event, centered within 50 m of the repository boundary, may result in transport of waste to the surface.

Using this approach, the probability of a volcanic eruption through the repository, given an eruption, can be approximated as

$$P[\text{eruption through repository} \mid \text{eruption centered at } x,y] = \begin{cases} 1, \text{if } (x,y) \in A_e \\ 0, \text{otherwise} \end{cases} \qquad (15)$$

where the effective area, A_e, is the area of the repository and the region about the repository within one conduit radius of the repository boundary (Geomatrix, 1996).

Other definitions of igneous events result in the need for more complex analyses of area affected because these events have length and orientation (Sheridan, 1992; Geomatrix, 1996). In these cases, probability density functions must be estimated for both the length and orientation of igneous events. Geomatrix (1996) gave the probability of an intrusive, igneous event centered on a given location intersecting the repository, which can be expressed as

$$P\left[L \geq l_r,\ \varphi_1 \leq \Phi \leq \varphi_2\right] = \int_{l_r}^{\infty} \int_{\Phi_1}^{\Phi_2} f_L(l) \cdot f_\Phi(\varphi)\ d\varphi dl \qquad (16)$$

where Φ is the azimuth of the igneous event with respect to north, with φ_1 and φ_2 representing the range of azimuths that would result in intersection with the repository, given an igneous event centered on x,y, a distance l_r from the repository boundary. The probability that the igneous event of half-length, L, will exceed l_r at an azimuth between φ_1 and φ_2 depends on the probability density functions $f_L(l)$ and $f_\Phi(\varphi)$ for igneous event half-length and azimuth, respectively.

This characterization of area affected by igneous events must be modified further depending on the type of event considered. Defining igneous events as volcanic vents or vent alignments may result in a probability estimate for volcanic disruption of the repository, if the frequency of vent formation along the alignment is included in the calculation. The length of the vent alignment is taken as the distance between the centers of the first and last volcanoes in the alignment. For example, the length of the Amargosa Aeromagnetic Anomaly A alignment of three vents is 4.0 km (Figure 7). The length of the Quaternary Crater Flat alignment of five vents is 11.2 km, based on the distance between southwest Little Cone and Northern Cone (Figure 5). Six vents occur along the 3.6-km Pliocene Crater Flat alignment. Average vent density along these alignments is on the order of 0.5-2.0 vents per km. This vent density suggests that, if an alignment defined by the distance between the first and last vents in the alignment intersects the repository, a vent will likely form within the repository boundary as a result of this intersection.

Uncertainty increases considerably when the functions $f_L(l)$ and $f_\Phi(\varphi)$ are introduced because these functions must be estimated from limited YMR geologic data. If the igneous event is defined as the development of a vent or vent alignment, mapped vent locations are useful in constraining the functions $f_\Phi(\varphi)$ and $f_L(l)$. Considering Plio-Quaternary volcanism in the YMR, six igneous events consist of the formation of isolated vents, and four igneous events resulted in the formation of vent alignments (Figure 13). Of these four vent alignments, two are less than 4 km long, the Pliocene Crater Flat vents and the Sleeping Butte vent pair. The Amargosa Aeromagnetic Anomaly A alignment is slightly longer than 4 km. The Quaternary Crater Flat alignment, one of the youngest and most important volcanic events in the YMR, is also at 11 km the longest alignment. Although these data provide an idea of the range of alignment lengths possible in the YMR, they are not sufficient to estimate a probability distribution for vent alignment lengths, $f_L(l)$.

In order to compensate for the lack of data within the YMR, analog information can be used. Draper, et al. (1994) note that approximately 30 percent of the vents in the San Francisco volcanic field form alignments. The remaining vents are isolated and appear to have formed during independent episodes of volcanic activity. This value appears comparable to the ratio of vent alignments to individual vents in the YMR. Data on vent alignment lengths from other volcanic fields suggests vent alignments may be considerably longer than the Quaternary Crater Flat alignment. For example, Connor, et al. (1992) identified vent alignments >20-km long in the Springerville volcanic field, Arizona. Vent alignments of comparable or greater length have been identified in the Michoacán-Guanajuato volcanic field, Mexico (Wadge and Cross, 1988; Connor, 1990), and the Pinacate volcanic field, Mexico (Lutz and Gutmann, 1995). Smith, et al. (1990) suggested alignments may be up to 20 km long, with a lower probability of 40-km-long alignments, based on mapping in the Lunar Crater, Reveille Range, and San Francisco volcanic fields. None of these authors, however, developed distributions for vent alignment lengths in these areas. Furthermore, it is not clear that the conditions for vent alignment formation and factors controlling vent alignment length are directly comparable between these different regions and the YMR. As a result, estimation of the distribution function for $f_L(l)$ for YMR vents and vent alignment formation is extremely uncertain.

However, given these caveats, the probability density function for the length of a new alignment is

$$f_L(l) = \frac{1}{2}\left[\delta(l) + \frac{1}{(l_{max} - l_{min})}\right] \quad (17)$$

where $\delta(l)$ is the delta function. By this definition, 50 percent of igneous events have zero length and only disrupt the repository if they fall within the effective area of the repository. The remaining 50 percent of igneous events form alignments that affect areas up to a distance l_{max} from the point *x,y*. This percentage assigned to zero-length igneous events is a source of uncertainty in probability estimates and is not well constrained by available data. The probability density function is construed to be a uniform distribution between l_{min} and l_{max} because the distribution of alignment lengths is so poorly known.

Using this definition of $f_L(l)$, probability estimates of intersection of the repository, given an event at *x,y*, depends on ($l_{max} - l_{min}$). Because l_{min} goes to 0, the analysis is most strongly dependent on the value of l_{max}. The value of l_{max} can be chosen as 5.6 km, taking the Quaternary Crater Flat alignment as the maximum alignment half-length. Given observations in other volcanic fields, however, l_{max} may be 10 km or more.

The distribution function for azimuth of alignments or dike zones, $f_\Phi(\varphi)$, is better constrained by the data on vent alignments, regional stress distribution, and the orientations of high-dilation tendency faults. Three of the alignments in the YMR trend 020° to 030°, perpendicular to the least principal horizontal compressional stress in the region, 028° (e.g., Morris, et al., 1996).

Under these circumstances, $f_\Phi(\varphi)$ may vary over a narrow range. For example,

$$f_\Phi(\varphi) = U\ [020^\circ,\ 035^\circ] \quad (18)$$

Alternatively, $f_\Phi(\varphi)$ near the repository may respond to the distribution of fault orientations (Figure 14) if ascending magmas tend to exploit faults as low-energy pathways to the surface (Conway, et al., 1997; Jolly and Sanderson, 1997).

Other definitions of igneous events attempt to capture the probability of igneous intrusions intersecting the repository boundary (Sheridan, 1992; Geomatrix, 1996). Igneous intrusions commonly form anastomosing networks at shallow levels in the crust, forming multiple dike segments at a given structural level (e.g., Gartner and Delaney, 1988). Consequently, a term may be added to Eq. (16) to account for the width of igneous events, such as the width of the dike swarm formed during igneous intrusion:

$$P\left[L \ge l_r,\ \varphi_1 \le \Phi \le \varphi_2,\ W \ge w_r\right] = \int_{l_r}^{\infty}\int_{\varphi_1}^{\varphi_2}\int_{w_r}^{\infty} f_L(l) \cdot f_\Phi(\varphi) \cdot f_W(w)\ dw\,d\varphi\,dl \quad (19)$$

where $f_W(w)$ is a probability density function describing the half-width of the igneous event, which may be a significant fraction of the half-length, and w_r is the shortest distance to the repository boundary perpendicular to the event azimuth, for a given azimuth and event length. Numerous individual dikes, dike segments, and sills may be located within this zone. Little is known about the distribution $f_W(w)$. In Pliocene Crater Flat, the half-width of the dike swarm appears to be on the order of 200 m. In contrast, Gartner and Delaney (1988) mapped dike zones up to 5 km wide (W = 2.5 km) in the San Rafael volcanic field (Figure 6).

Given the spatial density of these igneous features, it is conservative to consider intersection of the area defined by Eq. (19) with the effective repository area as resulting in igneous disruption of the site. This definition of an igneous event, however, does not necessarily result in direct transport of radioactive waste to the surface by erupting magma.

3.1.4.2 Summary

All probability models for volcanic disruption of the proposed repository rely on estimation of parameters to bound the temporal and spatial recurrence rates and magnitudes of igneous events. Ranges of these parameters adopted in the volcanic hazard analysis must be justified using geologic data and models. Estimation of the temporal recurrence rate relies on the frequency of past volcanic events in the YMR. These past recurrence rates indicate volcanism has persisted throughout the Pliocene and Quaternary at a low recurrence rate compared to many other Basin and Range volcanic fields. Therefore, such low temporal recurrence rates should be used to model probabilities. No evidence exists to indicate that basaltic volcanism has ceased in the YMR. Because the time elapsed since past volcanic eruptions within the YMR is short compared to common repose periods, the YMR should be considered a geologically active basaltic volcanic field, with recurrence rates greater than zero. Conversely, recurrence rates in the YMR are not as large as those in many other WGB volcanic fields, such as the Cima volcanic field where at least 30 volcanic eruptions have occurred since 1.2 Ma. Current evidence suggests that such an intense episode of volcanism is not likely in the YMR during the next 10,000 yr.

The temporal recurrence rate must be specified based on the definitions of igneous events. The current staff estimates for these recurrence rates are 2–12 v/m.y. for igneous events defined as individual mappable units or vents and 1–5 v/m.y. for vents and vent alignments. Staff concludes that new information presented in Wernicke, et al. (1998) and Earthfield Technology (1995) does not warrant a significant revision of recurrence rates used in NRC probability models. This new information, however, may affect probability models used by DOE (e.g., U.S. Department of Energy, 1998b) and as such will need to be addressed by DOE. The staff will continue to evaluate new information to determine the effects that it may have on estimated temporal recurrence rates. Temporal recurrence rate for igneous intrusions without volcanic eruptions is not estimated because data is not available to support such estimates. Based on analog data (Delaney and Gartner, 1997) a factor of two or greater is probably reasonable.

Spatial recurrence rate varies across the YMR because of vent clustering and the tendency for volcanism to recur within these clusters. For example, all Quaternary volcanism in the YMR occurs in proximity to Pliocene volcanoes. Estimations of spatial recurrence rate then must rely on patterns in past volcanic activity, which is done using kernel models. Spatial recurrence rates of igneous events at the repository or elsewhere on Yucca Mountain that are assumed to be at or near zero are not supported by existing data. Yet, spatial recurrence rates of zero or a slightly larger than zero regional background value are assumed at the repository in some models presented in Geomatrix (1996). Staff conclude that the distribution of sparse events does not provide an accurate basis to conclude that spatial recurrence rate within the repository boundary is zero or a low background value. Spatial analyses (e.g., Connor and Hill, 1995) indicate that the repository site is close to the edge of the Crater Flat cluster, within which most YMR Quaternary basaltic volcanism has occurred. A reasonably conservative model would, therefore, indicate that the spatial recurrence rate at the repository is greater than median spatial recurrence rates across the YMR.

Similarly, areas affected by igneous events must be described using parameter estimation, which will vary with the definition of igneous events. If igneous events are defined as individual mappable units and vents, then only those that erupt within the effective area of the repository significantly affect performance. Vent alignment lengths and orientations must be considered if igneous events are defined as vents and vent alignments. Vent alignment length is poorly constrained by available data, but its effect on probability is readily assessed using sensitivity studies. Alignment orientation is well constrained by the correlation between existing vent alignments and crustal stresses. Areas affected by igneous intrusions must be larger than areas affected by individual alignments, but the parameter distributions are poorly constrained.

3.1.5 Tectonic Models

3.1.5.1 Technical Basis

Probability models need to be consistent with tectonic models proposed for the YMR. Tectonic processes affect igneous processes across a large range of scales. Low recurrence-rate basaltic volcanic activity in the Basin and Range may occur where magmas are generated by decompression of fertile mantle during crustal extension (e.g., Bacon, 1982; McKenzie and Bickle, 1988). Magma ascent through the crust is enhanced by crustal structures produced by extension, leading to correlation between basaltic volcanism and structure across a range of scales, from the superposition of individual faults and vents to the occurrence of entire volcanic fields at the margins of extensional basins (Connor, 1990; Parsons and Thompson, 1991; Conway, et al., 1997). Volcanic hazard analysis of the proposed repository must quantify these often complex geological relationships.

The relationship between structure and volcanism has been used to suggest both higher and lower probabilities of volcanic disruption of the repository than are predicted using past spatio-temporal patterns in vent distribution alone (Connor and Hill, 1995). Smith, et al. (1990) suggested a narrow northeast-trending, structurally controlled source-zone of potential volcanism extends through the repository site, resulting in comparatively high probabilities of volcanic disruption. Alternatively, structure models that exclude the repository from volcanic source-zones result in comparably low probabilities. For example, Crowe and Perry (1989) proposed the north-northwest-trending CFVZ, with an eastern boundary located west of the repository site, effectively isolating the proposed repository. Thus, wide variation in probability estimates is a direct result of the varying ways in which these source zones have been drawn.

In the TSPA-VA, DOE uses source zones derived from Geomatrix (1996) to restrict the origin of an initiating dike to locations west of the proposed repository site (U.S. Department of Energy, 1998b). These source zones assume some fundamental geological differences occur between Crater Flat and Yucca Mountain, such that initiating igneous events are restricted to the Crater Flat source-zone. Although dikes of sufficient length can propagate from the source zone through the repository, this modeling approach biases, without sufficient basis, volcano locations away from the repository site such that the mean annual probability of volcanic disruption is $<10^{-8}$ (U.S. Department of Energy, 1998b; CRWMS M&O, 1998a). The same source-zone assumptions for many igneous events are used in subsequent probability models (CRWMS M&O, 2000b Framework), again concluding that the mean annual probability of volcanic disruption is $<10^{-8}$.

Although these source-zone examples often are referred to as structural models, none are defined by specific structural elements appearing on geologic maps or published subsurface structural interpretations (U.S. Nuclear Regulatory Commission, 1998c). Much of the confusion regarding volcanism source-zones could be resolved if the relationships between volcanism and structure are considered mechanistically and in light of mapped YMR structural features. In the following, current understanding of these relationships is discussed in terms of

- Regional tectonic models of Yucca Mountain and surrounding geologic features
- Mechanistic relationships between crustal extension and magma generation
- Local structural controls on magma ascent

3.1.5.1.1 Regional Tectonic Models

Yucca Mountain lies within the Basin and Range Province of the western North American Cordillera; a province characterized by spatially segregated regions of east-west extension between zones of northwest-trending, dextral strike-slip or oblique strike-slip faults. Coupled with the overall pattern of crustal extension and transtension are numerous small-volume volcanic fields (Figure 15). Within this tectonic framework, five viable tectonic models that describe the pattern of regional and local deformation around Yucca Mountain emerge from all those that have been proposed in the geologic literature during the past two decades (Stamatakos, et al., 1997b). These five models are

- Half-graben with deep detachment fault
- Half-graben with moderate depth detachment fault
- Elastic-viscous crust with planar faults with internal block deformation and ductile flow of middle crust
- Pull-apart basin (rhombochasm or sphenochasm)
- Amargosa shear or Amargosa Desert fault system

In a broad sense, these five models can be considered in two general categories of deformation. The first three are dominantly related to extensional deformation, and the latter are dominantly related to strike-slip deformation. Moreover, the five models are not mutually exclusive. Locally extensional-dominated deformation, within Crater Flat for example, can exist within a larger region of transtensional deformation related to a pull-apart basin.

In the deep detachment fault model (e.g., Ferrill, et al., 1996), the Crater Flat-Yucca Mountain faults are envisioned as soling into the Bare Mountain fault at the base of the seismogenic crust, at a depth of approximately 15 km (Figure 16a). The faults at Yucca Mountain accommodate strain within the hanging wall of the Bare Mountain fault. This model is dominantly extensional and compatible with a regional strike-slip system in which the Crater Flat-Yucca Mountain domain has largely dip-slip faulting, similar to a pull-apart basin. In addition, the model respects the geologic constraints on the timing of deformation (i.e., variable dips of fault blocks with growth of tuff strata across faults that were active during tuff deposition), as well as rollover in fault blocks. Restored cross sections, however, are more difficult to balance than with a moderate-depth detachment fault.

The moderate-depth detachment fault model (Young, et al., 1992; Ferrill, et al., 1995; Ofoegbu and Ferrill 1995) is similar to the deep detachment model, but the Crater Flat-Yucca Mountain faults sole into a detachment fault at 5–10 km depth (Figure 16b). The detachment then terminates against the deeper, larger Bare Mountain fault. The geometry of this model is the most reasonable for obtaining a balanced, restored cross-section of the upper crustal section.

Both shallow and moderately deep detachment models may influence basaltic magmatic activity in two ways. First, faults that sole into the detachment may serve as conduits for magma ascent in the shallow crust, if these faults provide relatively low-energy pathways to the surface (McDuffie, et al., 1994; Jolly and Sanderson, 1997). Second, dominantly extensional models result in large-scale density contrasts in the shallow crust. Relatively dense, PreCambrian and Paleozoic rocks dominate the upper crustal section west of the Bare Mountain fault. East of the Bare Mountain fault, extension results in the formation of a half-graben and the upper crustal section is dominated by less-dense tuffs and alluvium. This broad, density contrast may influence rates of partial melting, a topic discussed in Section 3.1.5.1.2.

Alternatively, Crater Flat-Yucca Mountain faults have been interpreted as planar to the ductile middle crust (Fridrich, 1998). This is an extension-dominant model; fault dips do not become more shallow with depth. This model, which serves as the conceptual basis for the United States Geological Survey boundary element model (Stamatakos, et al., 1997b), assumes the surface geometry of faults and fault blocks cannot be used to constrain deformation at depth. Internal fault-block deformation and ductile flow (and perhaps magma intrusion) at depth are assumed to compensate for variable fault-block dips, which otherwise would produce large triangular-shaped gaps in the subsurface.

The pull-apart basin model envisions Crater Flat as a pull-apart basin that formed in a releasing bend of a north-northwest-trending, regional strike-slip system (Minor, et al., 1997; Fridrich, 1998). The pull-apart basin is a half-graben with a well-defined western edge in the Bare Mountain fault, the diffuse set of Crater Flat-Yucca Mountain faults to the east, and an eastern edge in western Jackass Flats. The regional strike-slip system remains hypothetical, presumably buried beneath Amargosa Desert alluvium southeast of the southern end of the Bare Mountain fault. The pull-apart model explains the vertical axis rotation of the southern reaches of Crater Flat-Yucca Mountain (e.g., Hudson, et al., 1994) as crustal-scale block rotations within overall regional dextral shear. This shear is related to diffuse boundary interactions between the North American and Pacific plates. The model explains the north-northeast arcuate trend of Quaternary volcanic centers of Crater Flat as an alignment along a Reidel shear within the basin.

Fridrich (1998) has proposed two versions of this model. In the rhombochasm version of the pull-apart model, the basin-bounding, strike-slip fault trends north-northwest out of Crater Flat and is concealed beneath the Timber Mountain-Oasis Valley calderas. In the sphenochasm version, the northern extent of the bounding strike-slip fault is pinned at the northern end of Crater Flat. Strike-slip deformation increases south and east from the pin point. In response, the basin fans open to the south, and extension on basin bounding normal faults like the Bare Mountain fault increases southward (Scott, 1990; Stamatakos, et al., 1997a).

The Amargosa shear model is similar to the rhombochasm model, with Crater Flat representing a diffuse dextral shear-zone along a major north-northwest-trending crustal shear (e.g., Schweickert and Lahren, 1997). The shear zone extends northward along a hypothetical strike-slip fault extending north-northwest from Crater Flat beneath the Timber Mountain and Oasis Valley calderas. The lack of offset of these calderas is explained as diffuse detachment of the tuffs from underlying crust, in which offset is absorbed by horizontal faults within the tuff layers (Hardyman and Oldow, 1991). The southern extension of the shear links with the Stewart Valley-State Line fault. Total length of the fault and shear zones is greater than 250 km.

The Crater Flat shear zone includes the motion on faults within western Bare Mountain, the vertical axis rotation within southern Yucca Mountain, and the sites of volcanic activity in Crater Flat. The Quaternary cone alignment is believed to represent a Reidel shear oblique to the main shear axis. Based on a palinspastic reconstruction between southern Bare Mountain and the Striped Hills, this model calls for >30 km of right-lateral offset along the southern extension of this shear since 11.5 Ma (Schweickert and Lahren, 1997). This aspect of the model is suspect because of disparate exhumation ages for Bare Mountain and the Striped Hills, based on fission-track ages (Ferrill, et al., 1997) and paleomagnetic results (Stamatakos, et al., 1997c).

Strike-slip-dominated models have been used to infer an entirely different basis for distribution of volcanoes in the YMR other than purely extensional models. For example, Schweickert and Lahren (1997) envision a relatively uniform melt generation region beneath the YMR. In these circumstances, crustal structures such as Reidel shears in pull-apart basins allow magmas to ascend to the surface. Fridrich (1998) also proposed that tensional structures control the ascent of magma through the crust and that volcanism will be limited to areas where these tensional structures exist. Some source-zone probability models (e.g., Crowe and Perry, 1989) propose that Yucca Mountain lies outside of pull-apart basins, and, therefore, the probability of volcanism at Yucca Mountain is extremely low, compared with Crater Flat. As noted above, however, the strike-slip fault on the eastern edge of the pull-apart has not been mapped or identified. This lack of direct geologic evidence for a bounding fault on the east side of Crater Flat basin greatly reduces the confidence with which such source zones for basaltic volcanism can be drawn.

The amount of vertical axis rotation exhibited by Paintbrush and Timber Mountain formation tuffs is used by Fridrich, et al. (1999) to define rotational domains within the Crater Flat basin. They observe that <10.5 Ma basaltic volcanism is restricted to domains with more than 20° of vertical axis rotation. Although O'Leary (1996, p. 8–87) concludes that volcanic activity is not correlated with degree of vertical axis rotation, Fridrich, et al. (1999) and CRWMS M&O (1998a) use this degree of vertical axis rotation to define volcanic source-zones that restrict the proposed repository site from areas of future volcanism. As shown by Minor, et al. (1997) and Hudson, et al. (1994), vertical axis rotation began between 11.6–11.45 Ma during emplacement of Timber Mountain tuffs. Recent studies by Stamatakos and Ferrill (1996) and Stamatakos, et al. (2000) measured direction of remnant magnetization for 11.2 ± 0.1 Ma basalt in southern Crater Flat (Figure 5). These basalts overlie Timber Mountain tuffs that are rotated about 40°

clockwise (Hudson, et al., 1994; Minor, et al., 1997). In contrast, the 11.2 Ma Crater Flat basalts have a <10° counterclockwise rotation (i.e., Stamatakos and Ferrill, 1996; Stamatakos, et al., 2000), coincident with the minor vertical axis rotation measured in nearby 3.8 Ma Crater Flat basalt (i.e., Champion, 1991). Thus, the tectonic deformation that produced the significant vertical axis rotations occurred prior to eruption of basalt at 11.2 ± 0.1 Ma, and these basalts erupted in a different tectonic regime than was present during Timber Mountain tuff emplacement. The dikes emplaced near Solitario Canyon have a poorly constrained age (U.S. Nuclear Regulatory Commission, 1999) between 10.0 ± 0.4 and 11.7 ± 0.3 Ma but represent the same period of post-caldera volcanic activity as the 11.2 ± 0.1 Ma southern Crater Flat basalt (Perry, et al., 1998). The Solitario Canyon dikes were emplaced in Tiva Canyon Tuff, which has experienced no significant vertical axis rotation at the dike locations (Hudson, et al., 1994; Minor, et al., 1997). In contrast, many other areas to the south and west contained Tiva Canyon Tuff that had experienced up to 40° of vertical axis rotation. Thus, the degree of vertical axis rotation, which was formed prior to the basalt emplacement, did not define a structural domain that somehow controlled the location basaltic volcanic activity around 11 Ma (i.e., O'Leary, 1996). Although the locus of <12 Ma basaltic volcanic activity is in southwestern Crater Flat, coincident with the most likely zone of maximum crustal extension (e.g., Scott, 1990; Hudson, et al., 1994), volcanism clearly is not restricted to only areas of the highest crustal extension or vertical axis rotation. Models that define volcanic source-zones based on degree of vertical axis rotation (e.g., Fridrich, et al., 1999; CRWMS M&O, 1998a, 2000b) do not appear supported by available data.

Geophysical data for Yucca Mountain also provide some constraints on tectonic models and associated volcanic source zones. These data and associated models consistently show the Bare Mountain fault as the western boundary of the Crater Flat structural basin. Seismic reflection data in Brocher, et al. (1998) places the eastern bounding faults to Crater Flat structural basin significantly east of the Solitario Canyon fault, in the general vicinity of the Ghost Dance fault. Earthfield Technology (1995) provides a detailed evaluation of YMR aeromagnetic data within the limits shown in Figure 9. Magnetic basement maps in Earthfield Technology (1995) depict the eastern boundary of the Crater Flat structural basin in the area of the Paintbrush Canyon Fault. Minor, et al. (1997) use a pronounced gravity gradient east of Fortymile Wash to define the eastern boundary of the Crater Flat structural basin. Although the eastern boundary of the Crater Flat structural basin is often diffuse in these geophysical and tectonic models, these models clearly locate the proposed repository site within the Crater Flat structural basin. Consequently, volcanic source-zone models that localize volcanism away from the proposed repository site do not appear consistent with available geophysical data or tectonic models. Similarly, volcanic source-zone models that localize volcanism to narrowly defined zones intersecting the proposed repository site also do not appear consistent with available geophysical data or tectonic models.

Elements of the above tectonic models are not mutually exclusive. For example, predominately strike-slip deformation may have given way to predominantly extensional deformation as regional shear resulted in rotation of the direction of maximum horizontal compressional stress relative to fault planes. In light of these models, it is appropriate to consider mechanistic relationships between crustal extension in the YMR and basaltic magma generation. These relationships rely on a physical link between regional extension of the brittle crust and magma production deeper in the lithosphere.

3.1.5.1.2 Mechanistic Relationships Between Crustal Extension and Magma Generation

Crustal extension controls or strongly influences basaltic magmatism in the WGB (e.g., Leeman and Fitton, 1989; Lachenbruch and Morgan, 1990; Pedersen and Ro, 1992). Magmas that originate in WGB lithospheric mantle, including those of the YMR, were likely produced through decompression melting associated with extension (Farmer, et al., 1989; Hawkesworth, et al., 1995). Decompression melting is favored in zones of mantle lithosphere that have been previously enriched in incompatible elements, which enables melt formation at lower temperatures (e.g., McKenzie and Bickle, 1988). Based on mineralogical phase relationships and geochemical studies, decompression-induced lithospheric melting likely occurs at depths between 40–80 km (Takahashi and Kushiro, 1983; Rogers, et al., 1995). Extension and associated crustal deformation will produce local changes in lithostatic pressure at the base of the crust. Variations in lithostatic pressure produced through this extension may decompress enriched zones in lithospheric mantle sufficiently to partially melt and produce basaltic magma. Thus, lateral changes in lithostatic pressure across the YMR may control areas of future igneous activity.

Crustal extension has resulted in large density differences in the upper 5–6 km of the crust in the YMR due to the displacement of Paleozoic and PreCambrian rocks across the Bare Mountain fault, the formation of the Crater Flat basin, and subsequent deposition of tuff and alluvium in Crater Flat (Figure 17). The average density of a 5.6-km column of rock beneath Crater Flat and Bare Mountain can be calculated from this cross-section using average rock densities for the region (McKague, 1980; Howard, 1985). This difference in average density is 280 kg m^{-3}. Beneath this 5.6-km column, little density difference is expected because any faulting that occurs below 5.6 km does not juxtapose rocks of significantly different densities. Given lithostatic pressure as

$$P_L(z) = \int_0^z \rho(z) g \; dz \qquad (20)$$

where g is gravity (9.8 m/s^2), $\rho(z)$ is rock density at a given depth z, and z is the total depth (5.6 km), this density difference in the upper crust produces a lithostatic pressure difference between Bare Mountain and Crater Flat of approximately 15 MPa at a depth equivalent to the base of the Paleozoic section in Crater Flat. This lithostatic pressure estimate excludes topographic effects, because these effects attenuate rapidly with depth (Anderson, 1989).

Lateral changes in density at the surface, such as those produced by topographic variations or the development of a basin, attenuate with depth because of changes in the magnitudes of horizontal stresses relative to vertical stress as a function of depth. In this case, lithostatic pressure is best estimated as

$$P = -\frac{1}{3}\left(\sigma_{xx} + \sigma_{yy} + \sigma_{zz}\right) \qquad (21)$$

where σ_{xx}, σ_{yy}, and σ_{zz} are the orthogonal normal stresses.

Because of this attenuation, comparatively large-scale density variations are required to create lateral pressure changes in the mantle. Furthermore, lateral density contrast in the crust will cause lateral pressure changes in the mantle only if the Moho discontinuity is not deflected as a result of isostatic compensation (Figure 18). Isostatic compensation is not likely because the scale of features like Bare Mountain and Crater Flat are small compared to the scale of features normally compensated for by isostasy (Anderson, 1989). Existing geophysical data (Brocher, et al., 1996) support a flat Moho discontinuity in the YMR.

Bouguer gravity anomalies indicate that large-scale crustal density variations necessary to produce pressure variation in the mantle at >40 km occur in the YMR (Figure 19). The gravity map is dominated by large, negative anomalies produced by Timber Mountain-Oasis Valley calderas and a positive gravity anomaly associated with the Funeral Mountains. A north-trending area of largely negative gravity anomalies extends through Crater Flat and the Amargosa Desert.

These gravity data can be used to create an apparent crustal density map, following the methods of Gupta and Grant (1984), and to infer changes in apparent lithostatic pressure, ΔP_L, at comparatively shallow depths. Construction of the apparent density, or ΔP_L, map from the gravity data requires several assumptions:

- The gravity data must be on a regular grid. In this case, the gravity data were interpolated to a regular grid using a minimum tension bicubic-spline gridding algorithm.

- All density variation occurs due to lateral density variation between grid points. Density is taken to be constant between the surface and a depth, H, within each grid cell. Density variations in the Earth below H are not considered to contribute to the gravity anomalies.

- The method assumes a horizontal ground surface. The YMR gravity data have been reduced to a Bouguer anomaly, meaning density variations produced by topography and altitude effects have been removed from the gravity map. Using this data set results in lower density variation than expected, if topography is factored into the calculation. However, topographic effects have relatively short wavelengths, do not produce significant pressure differences at depths of magma generation, and, therefore, may be neglected.

Using the notation of Gupta and Grant (1984), the gravity anomaly at a point, $\Delta g(x,y)$, at the surface due to density variation at a point, $\Delta\rho(\xi,\eta,\zeta)$ beneath the surface, is

$$\Delta g(x,y,0) = G\left\{\frac{\partial}{\partial z}\int_{-\infty}^{\infty}\int_{-\infty}^{\infty}\int_{0}^{H}\frac{\Delta\rho(\xi,\eta)\,d\zeta\, d\eta\, d\xi}{\sqrt{(x-\xi)^2+(y-\eta)^2+(H-\zeta)^2}}\right\}_{z=0} \qquad (22)$$

where G is the universal gravitational constant. Note that, in this formulation, density does not vary as a function of depth. All density variation is lateral, and the amplitude of the gravity anomaly changes with depth of the anomalous mass only because of the change in distance from the mass anomaly to the gravity meter. Only the vertical component of the gravity anomaly is considered because this is measured by the gravity meter. Differentiating with respect to z gives

$$\Delta g(x,y,0) = G \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \int_{0}^{H} \frac{-\Delta\rho(\xi,\eta) d\zeta d\eta d\xi}{\left[(x-\xi)^2+(y-\eta)^2+\zeta^2\right]^{3/2}} \tag{23}$$

then integrating across depth

$$\Delta g(x,y,0) = G \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \frac{\Delta\rho(\xi,\eta) d\eta d\xi}{\sqrt{(x-\xi)^2+(y-\eta)^2}} - G \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \frac{\Delta\rho(\xi,\eta) d\eta d\xi}{\sqrt{(x-\xi)^2+(y-\eta)^2+H^2}} \tag{24}$$

which expresses the change in gravity in terms of the horizontal distance between the gravity meter and the density anomaly, and the average anomalous density averaged between the surface and depth H. Because all gravity variations are assumed to result from lateral variations in density, the relationship between gravity anomalies and apparent density anomalies can be expressed using a 2D Fourier transform of the gravity data. The 2D Fourier transform of the gravity field is given by

$$\Delta g(u,v) = \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \Delta g(x,y,0) \ \exp^{i(ux + vy)} \ dydx \tag{25}$$

where u and v are wave numbers. Gupta and Grant (1984) developed a simple filter to relate density and gravity in the wave number domain, based on the wavelengths of anomalies:

$$\Delta\rho(u,v) = \frac{1}{2\pi G} \times \frac{\varpi}{1-\exp^{-Z\varpi}} \times \Delta g(u,v) \tag{26}$$

where

$$\varpi = \sqrt{u^2 + v^2} \tag{27}$$

The inverse Fourier transform then yields apparent density in the spatial domain:

$$\Delta\rho(x,y) = \frac{1}{2\pi G} \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \frac{\varpi}{1-\exp^{-Z\varpi}} \Delta g(u,v) \ dvdu \tag{28}$$

The change in lithostatic pressure across the map region is then

$$\Delta P_L(x,y) = \Delta\rho(x,y) gH \qquad (29)$$

where g is now the average gravitational acceleration, 9.8 m s^{-1}, and H is the thickness of the crust within which all density changes are assumed to have occurred. Again, no significant density changes, in terms of overall change in lithostatic pressure, are assumed to occur at depths greater than H.

For H = 5000 m, $\Delta\rho(x,y)$ varies from approximately −100 to +240 kg m^{-3} across the YMR (Figure 20). The apparent density contrasts across the Bare Mountain fault in southern Crater Flat of 240–280 kg m^{-3} are in agreement with density contrasts obtained from the balanced cross-section and measured density values in the region (Figure 17). The most prominent feature of this map is the abrupt change in apparent density from high values west of the Bare Mountain fault to low values east of the Bare Mountain fault. Although this change is most abrupt adjacent to the Crater Flat basin, the apparent density map also reveals that this change persists south of Bare Mountain into the Amargosa Desert, and north of Bare Mountain. The apparent density map also shows that this change in density across the Bare Mountain fault is a long-wavelength feature. Apparent density values remain low east of the Bare Mountain fault for at least 50 km and high west of the Bare Mountain fault to the edge of the gravity map (Figure 20).

Because the magnitude of lateral pressure change will attenuate as a function of depth, only long-wavelength density variations in the crust will produce pressure changes in the mantle at depths of 40–80 km, the probable depth of magma generation in the YMR. The magnitude of pressure variations resulting from crustal density contrasts calculated across the Bare Mountain fault can be explored using finite element analysis. Based on a simplified geometric representation of the development of the basin, lateral pressure variations on the order of 7 MPa are expected to occur at depths of 40 km (Figure 18), attenuating to 2 MPa at a depth of 80 km, and « 1 MPa at 100 km. Mantle rocks at depths of 40–100 km are under average lithostatic pressures of 1000–3000 MPa. Thus, a change of 2–7 MPa across the density discontinuity represents a small fraction of the total pressure at that depth. This small difference reinforces the idea that extension and deformation of the magnitude observed in the YMR can only result in renewed magmatism if mantle rocks are already near their solidus (Figure 18).

Observations of the distribution of volcanoes in the YMR suggest that these small, lithostatic pressure differences are sufficient to generate basaltic melt. Plio-Quaternary volcanoes lie in the lower $\Delta P_L(x,y)$ areas east of the Bare Mountain fault, as expected if decreases in lithostatic pressure result in production of partial melts in the YMR. Nearly all of these volcanoes occur within the gravity low, which, in part, defines the Amargosa Gravity Trough (O'Leary, 1996) (Figure 19). Topographically, Lathrop Wells cinder cone lies outside Crater Flat but, based on gravity data, is within the larger north-trending basin and at the margin of the prominent basement low in southernmost Crater Flat. Aeromagnetic anomalies (Langenheim, et al., 1993) in the Amargosa Desert produced by buried Pliocene(?) basalts also lie within or at the margins of the southern extension of this basin. The easternmost of these buried basalts lies close to the north-trending gravity anomaly demarcating the eastern edge of the Amargosa Desert alluvial basin in this area.

These YMR volcanoes erupted in areas of lower $\Delta P_L(x,y)$ than expected if eruptions occurred randomly throughout the map area. In fact, only one Plio-Quaternary volcano erupted where $\Delta P_L(x,y)$ >+2 MPa, and this volcano, Aeromagnetic Anomaly E (U.S. Nuclear Regulatory Commission, 1999), erupted in a high gravity-gradient area along the southern projection of the Bare Mountain fault. These observations suggest that long-wavelength density differences in the YMR, dominated by displacement across the Bare Mountain fault and its apparent extension south into the Amargosa Desert, are sufficient to produce the pressure changes in the mantle that cause partial melting and volcanism.

This lithostatic pressure model suggests a correlation between the timing of extension and the timing of volcanism. Magma generated in response to extension, resulting in Quaternary volcanism within vent clusters formed by Miocene and Pliocene basaltic volcanism, occurred because mantle rocks beneath these regions were near their solidus and partially melted when comparatively small amounts of extension took place. A given rate of extension will result in the greatest rate of change in mantle pressure directly beneath the lateral change in crustal density, such as at the Bare Mountain fault. Thus, with continuing extension, mantle in the region of this inflection has the greatest opportunity of producing partial melts as a result of a given amount of crustal extension. Episodes of extension and basaltic volcanism may correlate temporally, because pressure variations in the mantle will likely equilibrate due to ductile flow over time. In other words, pressure changes in the mantle that result from crustal extension will be transitory.

Change in lithostatic pressure also affects magmatism, because magmas ascend by buoyant rise. The buoyancy forces acting on the magma are equivalent to the hydrostatic pressure gradient, given by Lister and Kerr (1991) as

$$P_h = \int_0^Z (\rho_{rock}(z) - \rho_{magma})\, g \, dz \qquad (30)$$

where ρ_{rock} and ρ_{magma} are densities of rock and magma, respectively, g is gravitational acceleration, and Z is the depth of magma generation. Rock density varies as a function of depth, most dramatically at the Moho. Because the density of magma is typically less than that of mantle, but greater than most crustal rocks, a level of neutral magma buoyancy exists in the crust. An isolated pod of magma above the level of neutral buoyancy sinks and a pod below the level of neutral buoyancy rises. Magmas fed by conduits respond to the integrated hydrostatic pressure along the conduit but also have flow characteristics that respond to the local hydrostatic pressure. Thus, dikes propagate laterally above the level of neutral buoyancy (Lister and Kerr, 1991). The level of neutral buoyancy is deeper in the crust beneath basins than beneath mountains. As Quaternary basalts in the YMR demonstrate, basalts do not stagnate in the alluvial basins as they rise through them because hydrostatic pressure is integrated over the depth from origination of the melt. Longer dikes and dike swarms, however, preferably form in these alluvial basins because of the basins' comparatively low lithostatic pressure. Thus, from the perspective of volcanic hazards analysis, understanding changes in lithostatic pressure across the region constrains areas of likely melt generation and areas of likely dike propagation above the level of neutral buoyancy.

3.1.5.1.3 Local Structural Controls on Magma Ascent

Observations in the YMR indicate a strong correlation between structure and volcanism. These observations include vent alignments (Smith, et al., 1990; Connor, et al., 1997) and cinder cones along faults (Section 3.1.4 and Connor, et al., 1997). These observations suggest that structural influences should be considered in PVHA of the proposed repository.

Basaltic magmas are transported from the mantle to higher levels in the crust or to the surface by igneous dikes. Propagating dikes, like other hydraulic fractures, typically form perpendicular to the least principal stress and parallel to the principal horizontal stress in extensional terrains (Stevens, 1911; Anderson, 1938).

Under some conditions, pre-existing faults or extension fractures serve as pathways for magma instead of propagating a new dike-fracture. Assuming that a pre-existing fault or extension fracture has no tensile strength, pre-existing fractures dilate (i.e., capture magma) if the fluid pressure exceeds the normal stress resolved on that fracture (Delaney, et al., 1986; Reches and Fink, 1988; Jolly and Sanderson, 1997). The likelihood of dilation and capture is controlled by the magnitude of the three principal stresses (σ_1, σ_2, σ_3), fluid pressure, and orientations of preexisting fractures in the *in situ* stress field.

The ability of any fault or fracture to dilate during magma injection is directly related to the normal stress acting across the fracture. Assuming cohesionless faults, the relative tendency for a fault of a given orientation to dilate in a given stress state (i.e., dilation tendency) can be expressed by comparing the normal stress acting across the fault with the differential stress (e.g., Morris, et al., 1996).

Dilation tendency of the fault is expressed as

$$T_d = \frac{(\sigma_1 - \sigma_n)}{(\sigma_1 - \sigma_3)} \qquad (31)$$

where σ_1 and σ_3 are the maximum and minimum compressional stresses, respectively, and σ_n is the normal stress acting across the fracture. Faults with T_d greater than some threshold value, such as 0.8, are considered to have a high dilation tendency (Morris, et al., 1996). A Schmidt plot of dilation tendency and fault poles indicates that, in the YMR, faults oriented 355–085° with dips >50° have a high dilation tendency (Figure 21).

In the YMR, σ_1 is vertical, σ_2 is horizontal and oriented 028°, and σ_3 is horizontal and oriented 298° (Morris, et al., 1996). The relative magnitudes of σ_1:σ_2:σ_3 are estimated to be 90:65:25. As a result of this stress pattern, steeply dipping, north-northeast-trending faults have a greater dilation tendency than faults of other orientations. Areas with higher concentrations of high dilation-tendency faults, therefore, are more likely to be the areas of volcanic activity. Cinder cone alignments form over prolonged periods of time if high dilation-tendency faults repeatedly serve as conduits for magma ascent (e.g., Conway, et al., 1997). McDuffie, et al. (1994) provide analytical results that show that the ability of a fault or fault zone to redirect ascending magma depends on the depth at which the dike intersects the fault and the dip of the fault zone. Only high-angle faults with dips greater than 40–50° are capable of dike capture at depths below

1 km. At depths of 10 km, faults dipping at angles less than 70° do not provide low-energy pathways to the surface, compared to vertical dike propagation.

Steeply dipping, high dilation-tendency faults in the YMR include many faults that bound the Yucca Mountain block, such as the Solitario Canyon and Ghost Dance faults. The Solitario Canyon fault adjacent to the repository site hosted dike injection at approximately 10.9 Ma. Moreover, the Solitario Canyon fault extends to the detachment fault at depths of 5–10 km (Figure 15). The distribution of faults with relatively high potentials for acting as magma conduits can be inferred from geologic mapping. In areas of alluvial cover, gravity and magnetic data provide the best indication of the distribution of these faults (e.g., Connor, et al., 1996).

3.1.5.2 Summary

Tectonic setting is important to consider in volcanic hazard analyses at several scales. On regional scales, crustal extension results in changes in pressure in the mantle and gives rise to partial melting. Extension also results in the formation of dip-slip fault systems, which serve as conduits for magma rise. On local scales and at shallow depths, individual dikes may propagate along faults that have high dilation tendencies and dike lengths may be controlled in part by local lithostatic pressure. Field investigations in the YMR have shown that all of these factors operate in the YMR, partially controlling the distribution and timing of basaltic volcanism.

Sufficient evidence exists to indicate basaltic volcanism in the western Great Basin (WGB) is linked to crustal deformation. Currently, several tectonic models are in use for the YMR, including detachment fault, simple horst and graben, Amargosa shear, and pull-apart models. Some commonality exists among these models with regard to basaltic volcanism. In particular, all of these models evaluate Crater Flat as an extensional half-graben, bounded on its western margin by the Bare Mountain fault. Although the eastern boundary of the Crater Flat structural basin is diffuse, most workers interpret this boundary east of the proposed repository site, usually between the Ghost Dance fault and the Fortymile Wash/Jackass Flat area. This structural basin appears to localize basaltic volcanism since about 12 Ma. Detachment fault, pull-apart, and Amargosa shear models all characterize the Bare Mountain fault as a major structure, transecting the brittle crust. The occurrence of the Bare Mountain fault can impact basaltic volcanism at several scales. On a regional scale, the Bare Mountain fault creates a substantial density contrast in the brittle crust. This density contrast causes changes in lithostatic pressure in the mantle that may induce partial melting. The Bare Mountain fault also may serve as a conduit for magma ascent through the brittle crust. The planar fault model is closer to a classical Basin and Range model of horst and graben formation (e.g., Stewart, 1971) than other tectonic models proposed for the YMR. However, this model shares elements with the other tectonic models in that the Bare Mountain fault is a major structure and Crater Flat basin is formed by extension (U.S. Nuclear Regulatory Commission, 1998c). Regardless of ultimate deformation mechanism, most of the tectonic models proposed to date include Yucca Mountain in the same structural domain as Crater Flat (Young, et al., 1992; Hudson, et al., 1994; Ferrill, et al., 1995; Ofoegbu and Ferrill, 1995; Schweikert and Lahren, 1997; Minor, et al., 1997; Stamatakos, et al., 1997a). Staff conclude that these models and available geophysical data reasonably demonstrate that the proposed repository site is located in the same structural domain that contains the <12 Ma basalts in Crater Flat basin. Although the locus of <12 Ma basaltic volcanic activity clearly lies southwest of the repository site, staff conclude that past patterns of igneous activity in Crater Flat basin accurately reflect the structural setting governing the likely locations of igneous activity during the next 10,000 yr. Probability models that restrict the location of future volcanism to sub-zones within the Crater Flat structural basin are not supported by available geophysical data or most structural models used in other aspects of the Yucca Mountain project (U.S. Nuclear Regulatory Commission, 1998c).

Results of a number of analyses indicate that incorporation of tectonic models into probability studies increases the probability of volcanic disruption of the proposed repository site compared to models that do not account for the tectonic setting of the site explicitly (Connor, et al., 1996; Hill, et al., 1996). This result primarily reflects the fact that Yucca Mountain is structurally part of the Crater Flat basin, with high dilation-tendency faults bounding and penetrating Yucca Mountain itself. Because of the presence of these structures, the lower limit on probability is represented by the nonhomogeneous Poisson models that do not incorporate structure. Probability models that incorporate tectonic features (e.g., the modified kernel model) are similar to some source-zone models in that the probability surface is elongate in a north-northwest direction, similar to the CFVZ proposed by Crowe and Perry (1989). The same tectonic features that enhance the probability of volcanism in Crater Flat, however, increase the probability of volcanism at Yucca Mountain, albeit to a lesser degree.

On local scales and at shallow depths, individual basaltic dikes may propagate along faults that have high dilation tendencies. Dike lengths may be, in part, controlled by local hydrostatic pressure. Field investigations in the YMR have shown that all of these factors may operate in the YMR, partially controlling the distribution and, possibly, the timing of basaltic volcanism. There is general agreement that volcano distribution is affected by local structural control. Dikes and vent alignments tend to be oriented northeast throughout the region, in response to horizontal stresses in the crust. Northeast trends have been accounted for in most analyses (e.g., Geomatrix, 1996; Smith, et al., 1990; Connor, et al., 1997).

3.1.6 Alternative Probability Models

3.1.6.1 Technical Basis

One of the difficulties inherent in the PVHA of the proposed repository is that the small number of volcanoes in the YMR makes it difficult to evaluate models quantitatively. Application of probability models in other volcanic fields (e.g., Condit and Connor, 1996) provides one method of evaluating probability models applied to the YMR. A second, equally important approach to model evaluation is to apply a range of models to estimate the probability of igneous events affecting the proposed repository and evaluate the sensitivity of probability estimates to bound the range of models. In the following, such a sensitivity analysis is performed for a range of models. The models differ primarily in how igneous events are defined and how more realistic, but often less well-constrained, geologic processes are included in the analysis. These probability models are based on

- Individual mappable eruptive units and vents
- Vents and vent alignments
- Vents and vent alignments with regional tectonic control
- Igneous intrusions

In the following sections, annual probabilities of igneous events are calculated and compared using these models and a range of parameters for recurrence rate and area affected by volcanism.

3.1.6.1.1 Individual Mappable Eruptive Units and Vents

Individual mappable eruptive units and vents were used by Connor and Hill (1993, 1995) to estimate the probability of volcanic eruptions at the site. This definition of igneous events involves the fewest assumptions about volcanism, resulting in a straightforward sensitivity analysis.

Assuming that the probability of more than one event in a given year is small, the annual probability of volcanic eruptions within the repository boundary is given by

$$P[\text{volcanic eruptions within repository boundary}] = 1 - \exp[-\lambda_r \, \lambda_t \, A_e] \tag{32}$$

where λ_t is the annual regional recurrence rate of volcanic vent formation, A_e is the effective repository area (Geomatrix, 1996), and λ_r is the spatial recurrence rate of volcanic eruptions at the repository, given a volcanic event in the region. Using a Gaussian kernel

$$\lambda_r(x,y) = \frac{1}{2\pi h^2 N} \sum_{v=1}^{N} \exp\left\{-\frac{1}{2}\left[\left(\frac{x-x_v}{h}\right)^2 + \left(\frac{y-y_v}{h}\right)^2\right]\right\} \tag{33}$$

where x,y is a Cartesian coordinate within the repository boundary, x_v,y_v is the coordinate of the center of an igneous event, N is the number of such igneous events, *and* h is a smoothing parameter (Section 3.1.4.3). For the following calculations, x,y is 548500, 4078500 and x_v,y_v are in Universal Transverse Mercator coordinates (U.S. Nuclear Regulatory Commission, 1999). Based on the analysis in Section 3.1.4.3, a smoothing parameter, $h \geq 5$ km, is appropriate for the Gaussian kernel. An effective repository area of 5.49 km is used in this analysis, based on the current repository design (Figure 5) and a 50-m buffer zone about the repository perimeter. The number of igneous events, N, depends on whether Pliocene and Quaternary or only Quaternary volcanoes are considered in the probability estimate.

Eight igneous events have occurred in the YMR during the Quaternary, if these events are defined as individual mappable eruptive units and vents. Connor and Hill (1995) used this definition for igneous events and varied recurrence rates between 5–10 v/m.y. Here, we model a range of 2–12 v/m.y. A recurrence rate >12 v/m.y. would signal a marked increase in activity compared to other WGB volcanic fields. Recurrence rates in the Cima volcanic field, California, which is one of the most active basaltic volcanic fields in the WGB, are on the order of 30 v/m.y. (Turrin, et al., 1985). Comparable rates of basaltic volcanism have not occurred during the Plio-Quaternary in the YMR, with the possible exception of in the Funeral Formation. Rates of less than 2 v/m.y. would signal a marked decrease in magmatism in the YMR. No evidence currently available suggests such a decrease is likely. Therefore, the assumption that such a decrease in regional recurrence rate will occur can not be supported for the volcanic hazard analysis.

Estimated probabilities using this model are sensitive to temporal recurrence rate of igneous events in the YMR, λ_t, and choice of h in the calculation of $\lambda_r\,(x,y)$ (Figure 22). Based on these parameters, the annual probability of volcanic eruptions within the repository boundary is between 0.5×10^{-8} and 3.5×10^{-8}. Probabilities are slightly higher if the distribution of Quaternary volcanoes is considered in estimation of λ_r, rather than the distribution of Plio-Quaternary volcanoes, because Quaternary volcanoes are, on average, located closer to the repository site. These values are quite close to those calculated by Connor and Hill (1995) using Epanechnikov kernel and nearest-neighbor estimators of spatial and spatio-temporal recurrence rate. Connor and Hill (1995) used A_e = 8 km^2 and estimated annual probabilities of volcanic disruption of the site between 1×10^{-8} and 5×10^{-8}.

3.1.6.1.2 Vent Alignments

If igneous events are defined as vents and vent alignments, probability of volcanic eruptions within the repository boundary incorporates distance and direction of an igneous event centered at a point, x,y, from the repository boundary. The probability of an igneous event centered at x,y is given by

$$P_{x,y}\ [\text{ igneous event at } x,y\ ] = 1 - \exp\left(-\lambda_t \lambda_r \Delta x \Delta y\right) \tag{34}$$

where λ_t is the regional recurrence rate and λ_r is the spatial recurrence rate at point x,y, calculated using the Gaussian kernel [Eq. (33)]. In practice, λ_r is calculated on a grid of points with map extent X,Y and grid spacing Δx, Δy. This probability is then weighted by the probability that an igneous event centered at $x,y,$ or occurring within Δx, Δy will result in a volcanic eruption within the repository boundary. For vent alignments in the YMR, the spacing of vents along the alignments is small compared to the size of the repository (Section 3.1.3.2). Vent alignment length is defined as the distance between the centers of the first and last vents on the alignment. Therefore, the probability that an igneous event centered at x,y will result in vent alignment intersection with the repository boundary and subsequent volcanic eruption within the repository boundary is

$$P_{lr}[\text{volcanic eruptions within repository boundary} \mid \text{igneous event at } x,y] \\ = \begin{cases} 1, & x,y \in A_e \\ \dfrac{1}{2}\left[\dfrac{l_{\max} - l_r}{l_{\max} - l_{\min}}\right], & l_{\min} \le l_r \le l_{\max} \\ 0, & l_r > l_{\max} \end{cases} \tag{35}$$

where $l_{\min}$ and $l_{\max}$ are the minimum and maximum alignment half-lengths, respectively, and l_r is the distance from x,y to the nearest repository boundary along the direction of the alignment. For this analysis, vent alignments are assumed to be oriented 028°, perpendicular to the direction of minimum compressional stress in the YMR. Experimentation indicates that choosing a range of values of alignment orientation between 020° and 035° has a negligible effect on probabilities of volcanic eruptions within the repository boundary. Probabilities are sensitive to

l_{max}, which is varied over a range of values in the following analysis, but are not sensitive to the selection of l_{min}, which for the following calculations is 100 m. As indicated in Eq. (35), 50 percent of all igneous vents are not part of vent alignments in this model. The probability of volcanic eruptions within the repository boundary is then

$$P[\text{volcanic eruptions within the repository boundary}] = \sum_{i=1}^{X} \sum_{j=1}^{Y} P_{x,y}(x_i,y_j) \cdot P_{lr}(x_i,y_j) \tag{36}$$

where x_i, y_j are on a rectangular grid of extent X,Y and grid spacing Δx, Δy.

Annual probability of volcanic eruptions within the repository boundary were calculated using 5200 m $\leq l_{max} \leq$ 10,200 m, and h = 5 and 7 km (Figure 23). Based on nearest-neighbor vent and vent alignment distances in the YMR, $h \geq$ 7 km is reasonably conservative (Figure 10). Using three Quaternary igneous events (Lathrop Wells, Quaternary Crater Flat, Sleeping Butte), results in annual probabilities of volcanic eruptions within the repository boundary between 1×10^{-8} and 3×10^{-8}, assuming a regional recurrence rate of 3 v/m.y. A rate of 5 v/m.y. results in annual probabilities of 6×10^{-8}.

3.1.6.1.3 Vent Alignments With Tectonic Control

For a more complete analysis, the previous probability estimates should be modified to incorporate additional geologic controls on volcanism (e.g., Connor, et al., 2000). Tectonism in the YMR has led to regional variations in crustal density that may cause variation in rates of partial melting throughout the YMR (Section 3.1.4.1). These variations are most apparent across the Bare Mountain fault. Plio-Quaternary basaltic volcanism clusters east of this fault, in areas of anomalously low crustal density. In contrast, basaltic volcanism since the mid-Miocene is apparently absent west of the Bare Mountain fault and its southern extension into the Amargosa Desert. Standard Gaussian kernel functions do not take into account these geologic details. As a result, the standard Gaussian kernel [i.e., Eq. (33)] is too simple and overestimates probabilities of volcanic eruptions in some areas (e.g., Bare Mountain) and underestimates probabilities elsewhere in the YMR.

The standard Gaussian kernel model developed previously was modified by developing a weighting function that accounts for crustal density. The model for basaltic volcanism in extensional environments developed in Section 3.1.5.1 and Connor, et al. (2000) relates lithostatic pressure gradients in the mantle to regional changes in crustal density caused by extension. As illustrated in Figure 18, partial melting occurs where partial melting had occurred previously and close to active graben-bounding faults where slip in the crust causes the greatest pressure change in the mantle.

Pressure change in the mantle is inferred conceptually from simple numerical models of mantle stresses (Figure 18). The weighting function can be estimated from the frequency of volcanic eruptions as a function of crustal density. The distribution of this function, $f_T(x,y)$, was defined based on average crustal densities in the upper 5 km of the crust at the locations of existing volcanoes, derived from application of the density filter to the gravity data set (Figure 24). The Gaussian kernel was then modified to estimate the recurrence rate of volcanism at x,y:

$$K_g(x_i,\ y_j) = \exp\left\{-\frac{1}{2}\left[\left(\frac{x_i - x_v}{h}\right)^2 + \left(\frac{y_j - y_v}{h}\right)^2\right]\right\} \tag{37}$$

$$Q_v = \frac{\sum_{i=1}^{X}\sum_{j=1}^{Y} K_g(x_i,\ y_j)}{\sum_{i=1}^{X}\sum_{j=1}^{Y} f_T(x_i,\ y_j)\cdot K_g(x_i,\ y_j)} \tag{38}$$

$$\lambda_r(x,y) = \frac{1}{2\pi h^2 N}\sum_{v=1}^{N} Q_v\ f_T(x,y)\ K_g(x,y) \tag{39}$$

Introduction of the ratio Q_v assures that the integral of the modified Gaussian kernel for a single volcano for a large map extent *X,Y* relative to the smoothing parameter, *h*, will be unity [Eq. (5)]. The probabilities, however, are redistributed based on crustal density variations in the vicinity of the volcano.

Comparison of the modified and standard kernels was made by contouring $\lambda_r(x,y)$ throughout the YMR, using the distribution of Quaternary vents and vent alignments and h = 9000 m. As previously, N = 3 in this model, defined by Quaternary Crater Flat, Lathrop Wells, and Sleeping Butte as the three Quaternary igneous events. In Figure 25, $\lambda_r(x,y)$ is contoured across the map region using Eq. (33). Given an igneous event in the region, there is a 68-percent chance that the igneous event will occur within this map area. The Sleeping Butte alignment lies north-northwest of the mapped region (see Figure 2). Larger values of $\lambda_r(x,y)$ indicate areas where igneous events are most likely centered. The largest values occur in southern Crater Flat because of the proximity of Lathrop Wells and the Quaternary Crater Flat alignment. In this area, $\lambda_r(x,y)$ varies between 8×10^{-4} volcanic events per square kilometer (v/km^2) and 2×10^{-4} v/km^2.

Figure 26 is based on the modified kernel [Eqs. (37)–(39)] using the same parameters as used in the standard kernel calculation (N = 3, h = 9000 m), but weighting the kernel using crustal densities derived using Eqs. (22) to (29). Use of the modified kernel reduces the area of the $\lambda_r(x,y)$ surface at, for example, the 2×10^{-4} v/km^2 contour and increases the amplitude of the surface. The $\lambda_r(x,y)$ surface also becomes asymmetric as a result of application of the modified kernel function. Values of $\lambda_r(x,y)$ are greatest in southern Crater Flat, exceeding 1.2×10^{-3} v/km^2, and decrease abruptly near the Bare Mountain fault. Probability values decrease less abruptly on the eastern boundary of Crater Flat because crustal densities change less rapidly on the eastern edge of the basin. This more gradual change in $\lambda_r(x,y)$ on the eastern edge of the basin is consistent with the proposed model linking crustal extension and basaltic volcanism (Figure 18).

The annual probability of volcanic eruptions within the repository boundary increases when the modified kernel function is used. Annual probability of volcanic eruptions within the repository boundary was calculated using 5200 m $\leq l_{max} \leq$ 10,200 m, and h = 7 km (Figure 27). Using the three Quaternary igneous events (Lathrop Wells, Quaternary Crater Flat, Sleeping Butte) results in annual probabilities of volcanic eruptions within the repository boundary between 3×10^{-8} and 5.5×10^{-8}, assuming a regional recurrence rate of 3 v/m.y. Including Pliocene volcanoes in the estimation of $\lambda_r(x,y)$ decreases the annual probability at the repository because many Pliocene volcanoes are located in the Amargosa Desert. Annual probabilities based on the modified kernel distribution and Plio-Quaternary volcanoes vary between 1.5×10^{-8} and 3×10^{-8}, comparable to the annual probabilities estimated using the standard kernel and the distribution of Quaternary vents and vent alignments. The regional recurrence rate of vent and vent alignment formation is poorly constrained in the YMR. Varying regional recurrence rate of igneous events between 1 and 5 v/m.y. results in nearly one order of magnitude variation in the annual probability of volcanic eruptions within the repository boundary. Using the modified kernel model, h = 7 km, and 5200 m $\leq l_{max} \leq$ 10,200 m, the annual probability of volcanic eruptions within the repository varies between 1×10^{-8} and 9×10^{-8} (Figure 28; Connor, et al., 2000).

3.1.6.1.4 Igneous Intrusions

The probability of igneous intrusions, such as dike swarms, intersecting the repository is greater than the probability of volcanic eruptions within the repository, because igneous intrusions must have greater areas than vent alignments and most likely occur with greater frequency. All alignments have associated intrusions but not all intrusions produce vent alignments. The recurrence rate of igneous intrusions and their geometry, however, are so poorly constrained by available data that these parameters are not estimated. Based on analogy with the San Rafael volcanic field (Delaney and Gartner, 1997), probabilities of igneous intrusion into the repository boundary may be two to five times the probability of volcanic eruptions within the repository boundary. While such a value is speculative it does provide a basis for development of an interim probability value for igneous intrusion intersecting the repository.

3.1.6.2 Summary

Annual probability of volcanic eruptions within the repository boundary varies between 10^{-8} to 10^{-7} based on a range of models (Connor, et al., 2000). This range accounts for varying definitions of igneous events and uncertainty in parameter distributions used to estimate probability. As discussed in Section 3.1.4.1.1 of this report, staff conclude that the past patterns of volcanic activity accurately represent volcanic recurrence rates for use in YMR probability models. Staff conclude that strain-rate data presented in Wernicke, et al. (1998) or the anomalies identified in Earthfield Technology (1995) do not provide a reasonable technical basis to conclude the volcanic recurrence rates used herein have been underestimated significantly for the proposed repository site. Additional basaltic centers identified in Magsino, et al. (1998) and Connor, et al. (2000) also will not affect significantly an annual probability range of 10^{-8} to 10^{-7}. This, however, is not the case with the probability models provided within the Viability Assessment (U.S. Department of Energy, 1998b) or supporting documents (CRWMS M&O, 2000b). The event counts used in these various models would need to be re-evaluated, based on the new data. As a strong reliance has been placed on the Probabilistic Volcanic Hazard Assessment (Geomatrix, 1996), it is impossible to say, without discussion with the individual panel members, what this information would do to the event counts and factors for undetected events used by the various panel members. At best, the staff consider the recurrence rates, and hence the probability values, as lying at the low end of the range of acceptable values.

Annual probabilities are generally between 1×10^{-8} and 3×10^{-8} for igneous events defined as individual mappable units and vents. This definition of igneous events requires the fewest assumptions about underlying parameter distributions but also neglects some features of vent distribution that are important in the YMR. In particular, the formation of vent alignments is not accounted for in this model. Defining igneous events as vents and vent alignments results in a similar range of probability estimates for the annual probability of volcanic eruptions within the repository boundary, 1×10^{-8} to 6×10^{-8}. Although recurrence rates are lower using this definition of igneous events, the area affected by individual events is greater. The distribution of alignment length and regional recurrence rate of these igneous events introduces the greatest uncertainties into these probability models. Incorporating regional crustal density variation into this model results in a model more closely linked to geologic processes. Based on the crustal density models and similar models presented previously (Hill, et al., 1996; Connor, et al., 1996), the annual probability of volcanic eruptions within the repository boundary is between 1×10^{-8} and 9×10^{-8}. Probabilities of intersection of igneous intrusions with the repository are likely higher but cannot be confidently estimated from available geologic data. As a value is needed for use in performance assessment, NRC will assume the rate is a factor of between 2 to 5 higher than for volcanic disruption. Finally, it is noted that this range of probability values, 10^{-8} to 10^{-7}, arises from the application of a variety of models and a range of parameter distributions. Nothing in the above analysis suggests that this range of probabilities has central tendency, that the mean or median of this range of probabilities is significant, or that high or low values in this range are more or less likely. This situation arises because, at least at the current time, it is not feasible to develop an objective basis for assigning likelihood to individual models, due to both lack of data and uncertainty in our understanding of the process. For the purpose of performance assessment, the NRC will assume the value of 10^{-7} for volcanic disruption of the proposed repository site. As the NRC recognizes the potential effect on probabilities that the new information discussed above could produce, based on the models used in this report, the NRC sees no present basis for changing this value and considers that the new information further justifies the use of the 10^{-7} value.

The WGB, which includes Yucca Mountain, is a magmatic province characterized by Quaternary basaltic volcanism (Fitton, et al., 1991). At least 211 basaltic volcanoes <2 Ma occur in the 82,000 km^2 region defined by Amboy volcano, the Big Pine volcanic field, and the Lunar Crater volcanic field (Figure 1; Luedke and Smith, 1981; Connor and Hill, 1994). Assuming that volcanism is randomly distributed throughout this source-zone (cf. Crowe, et al., 1995; Geomatrix, 1996), volcano recurrence rates are 1.3×10^{-9} yr^{-1} km^{-2}. The annual probability of volcanic disruption of any 5-km^2 area (i.e., repository area) in this source zone is thus 6×10^{-9}. This analysis overlooks the fact that volcanoes cluster within the WGB (Figure 1). The YMR, however, constitutes one of the volcano clusters within the WGB (Connor and Hill, 1995), within which probability should be higher than expected, based on a uniform random model. An annual probability of 6×10^{-9} appears a reasonable and general measure of background volcano occurrence for any 5-km^2 area within the WGB, including the Yucca Mountain repository site. Models that propose an annual probability of volcano formation at the proposed repository site of less than 6×10^{-9}, thus, do not appear to be reasonable, based on geologic data.

The likely regional background rate for basaltic intrusions is necessarily higher than that of single volcanoes, due to the larger area affected by a shallow basaltic dike. Using conditions appropriate for the Yucca Mountain repository site, the regional probability of a shallow basaltic intrusion can be assessed by sampling a uniform random distribution of dike half-length between 0.1–4 km and trending 28° from north. The annual probability of igneous disruption of any 5-km^2 area in the WGB is then 1.7×10^{-8}. This simple calculation does not consider the possibility of unmapped shallow dikes that were emplaced without an associated volcanic eruption or the presence of misdated Quaternary cinder cones in the WGB. Models that propose an annual probability of igneous dike intersection with the proposed repository site of less than 1.7×10^{-8} do not appear to be reasonably supported.

3.1.7 Probability Model Uncertainty

3.1.7.1 Technical Basis

A deterministic approach evaluates uncertainty by bounding model parameters. Parameter values are generally selected such that overall risk is not underestimated. This approach results in a single, straightforward value that bounds performance but does not provide any quantitative information on the uncertainty associated with this value (U.S. Nuclear Regulatory Commission, 2000). Detailed documentation and justification for parameter values used in this approach are required in order to determine the appropriate level of conservatism needed to represent the range of data.

A probabilistic approach provides a distribution of model results, which, in turn, provides a quantitative measure of uncertainty. This approach is more objective than a deterministic approach in that a level of conservatism is not implicitly required. The range of parameter values must be reasonable, and appropriate sampling methods must be used in the analysis (U.S. Nuclear Regulatory Commission, 2000). The mean value of a probabilistic analysis is generally used to determine compliance with the performance objective U.S. Nuclear Regulatory Commission, 2000). For low-level waste licensing, NRC staff also recommended that the 95th percentile of the performance distribution be less than a given value to demonstrate compliance (U.S. Nuclear Regulatory Commission, 2000). Because NRC is using a single value in performance assessment for volcanic probability, it is further justification of the use of the value of 10^{-7}.

Uncertainty associated with any probability model consists of two components that measure precision and accuracy. Precision is also referred to as “parameter uncertainty,” whereas, accuracy often reflects “model uncertainty” (U.S. Nuclear Regulatory Commission, 2000). Of the range of probability models proposed for the YMR, only the spatio-temporal nonhomogeneous models of Connor and Hill (1995) have been evaluated for model accuracy (Condit and Connor, 1996). This initial evaluation demonstrates that these probability models reasonably estimate the locations of basaltic volcanoes in the Springerville volcanic field when basalt petrogenesis remains relatively constant. These models are unsuccessful in estimating the future locations of basaltic volcanoes when the magmatic system undergoes abrupt and large shifts in petrogenesis (Condit and Connor, 1996). The YMR has not undergone similar magnitude petrogenetic shifts since about 5 Ma (e.g., Crowe, et al., 1986), thus, these probability models should be reasonably accurate when applied to the YMR system.

3.1.7.2 Summary

Based on the range of work currently available, the probability of igneous events at the proposed repository site can be described by single values, mean values of various distributions, entire probability distributions, or bounds on probability distributions. Any of these approaches may be used based on current NRC regulations. Regardless of the value(s) utilized, the methods used to derive the values must be justified, and the data used to derive the values must be clearly presented. In addition, probability models used in licensing must be shown to reasonably forecast the timing and location of future igneous events.

3.1.8 Expert Elicitation

3.1.8.1 Technical Basis

As summarized in U.S. Nuclear Regulatory Commission (1996), the NRC expects that subjective judgments of groups of experts will be used by DOE to assess issues related to overall performance of the proposed high-level radioactive waste repository site at Yucca Mountain. NRC has traditionally accepted expert judgment as part of a license application to supplement other sources of scientific and technical data. Expert elicitation is commonly used when

- Empirical data are not reasonably obtainable or analyses are not practical to perform.
- Uncertainties are large and significant to a demonstration of compliance.
- More than one conceptual model can explain, and be consistent with, the available data.
- Technical judgments are required to assess whether bounding assumptions or calculations are appropriately conservative.

U.S. Nuclear Regulatory Commission (1996) also summarize a series of technical positions and procedures concerning the use of expert elicitation in demonstrating compliance with geologic repository disposal regulations. These procedures emphasize the need for detailed documentation during the elicitation and for transparency in the aggregation of multiple expert’s judgments. An elicitation also should provide a means to evaluate new data that may arise between completion of the elicitation and submittal of licensing documents (U.S. Nuclear Regulatory Commission, 1996).

DOE used expert judgement to arrive at a probability value for igneous activity at the repository site (Geomatrix, 1996). Although the report generally followed the NRC Branch Technical Position (BTP) regarding expert elicitation (U.S. Nuclear Regulatory Commission, 1996), several areas of weakness in the elicitation procedure were noted in the September 1996 Appendix 7 meeting with DOE:

- Criteria and procedures for incorporating new data into the existing elicitation need to be established and published.
- Central issues need to be deconvoluted as much as possible, so that standard definitions of terms can be used consistently throughout the elicitation.

- Greater balance is needed on the panel to encompass a wider range of viewpoints, along with more thorough documentation of the selection processes and potential conflicts of interest for panel members.

- Intermediate judgments of the experts after the elicitation and any changes of rationales need to be documented.

Following the Appendix 7 meeting, NRC concluded that the elicitation (Geomatrix, 1996) is generally consistent with the BTP regarding the conduct of an expert elicitation. NRC will, thus, give the elicitation the appropriate level of consideration in the review of licensing documents (Bell, 1997).

Staff have performed a technical review of the PVHA elicitation report (Geomatrix, 1996) and, as explained in previous sections of this report, have several technical concerns regarding the PVHA results and their application in the Yucca Mountain program. The most significant concern is that many of the models in the PVHA are critically dependent on the definition of volcanic source-zones. Many of the source-zone models bypass the proposed repository site due to a lack of previous igneous activity at the site (Geomatrix, 1996). Although some geological data appear to suggest such division, critical analyses reveal that these apparent divisions are only manifestations of surficial features and not important to deeper structural control of volcanism (e.g., Stamatakos, et al., 1997b). In addition, larger-scale geologic features that commonly affect the localization of basaltic igneous activity are remarkably similar between the proposed repository site and the locations of past igneous activity. Based on these geologic relationships, staff conclude that volcanic source-zones that fail to include the proposed repository site are not reasonably conservative.

According to Geomatrix (1996) mean annual probability of repository disruption is 1.5×10^{-8} yr^{-1}. This is, however, a combined probability for both volcanic and intrusive igneous events. Utilizing the source zone models that preclude volcanoes from forming at the repository site, as was done repeatedly in Geomatrix (1996), requires that the actual probability of volcanic disruption based on this methodology is necessarily lower than 1.5×10^{-8} yr^{-1}. A rough estimate is that the mean PVHA probability for volcanic disruption may be an order of magnitude lower than the combined probability for all classes of igneous events. In order to use probability estimates in performance assessment they must, in some way, be separated into volcanic and intrusive events. In TSPA-VA (U.S. Department of Energy, 1998b), the igneous event probabilities from the PVHA elicitation were erroneously referred to as probabilities of volcanic disruption. In order to derive probabilities of volcanic disruption of the proposed repository site from the igneous event probabilities in the PVHA (Geomatrix, 1996), CRWMS M&O (1998a) used an average dike intersection probability of 1.5×10^{-8} yr^{-1} from Geomatrix (1996). Dikes were assumed to originate in a volcanic source zone that did not include the proposed repository site, thus, every dike in CRWMS M&O (1998a) extended beyond the repository boundaries. CRWMS M&O (1998a) then assumed 1–5 volcanic vents could localize randomly along the dike, resulting in 0–4 vents potentially localizing within the repository footprint. This method resulted in an average annual probability of volcanic disruption around 6×10^{-9} in CRWMS M&O (1998a). As noted in section 3.1.6.4 of this report, staff considers an annual probability of 6×10^{-9} as representative of background hazard rates for randomized volcanism throughout the entire WGB region and not representative of the long history of recurring basaltic volcanism in the YMR.

Although CRWMS M&O (1998a) developed a new methodology to interpret the results of Geomatrix (1996), significant amounts of new information developed subsequent to the PVHA elicitation was not addressed or incorporated into these interpretations. Many of the probability models in Geomatrix (1996) used volcanic source-zones, defined in part by panel members understanding of the structural setting of the Yucca Mountain area. Recent structural studies by Hudson, et al. (1994), Langenheim and Ponce (1995), O'Leary (1996), U.S. Nuclear Regulatory Commission (1997), and Minor, et al. (1997), in addition to new geophysical information in Earthfield Technology (1995), U.S. Nuclear Regulatory Commission (1997), and Brocher, et al. (1998) provide technical bases to conclude the proposed repository site is within the same structural setting (i.e., volcanic source-zone) as basaltic volcanoes located in the Crater Flat area. In addition, data and analyses presented in Earthfield Technology (1995), U.S. Nuclear Regulatory Commission (1997), Wernicke, et al. (1998) and Magsino, et al. (1998) question the validity of recurrence rate estimates used in Geomatrix (1996). This new information was not utilized in CRWMS M&O (1998a) and resulting conclusions in U.S. Department of Energy (1998b).

3.1.8.2 Summary

There are no generally accepted methodologies for calculating the probabilities of future igneous activity in distributed volcanic fields over periods of 10,000 yr. In addition, more than one conceptual model can be applied to this problem, resulting in a wide range of probability values. DOE is using expert elicitation (Geomatrix, 1996) to evaluate a range of probability models, estimate uncertainties in model results due to reasonable variations in model parameters, and determine a probability distribution for use in performance assessment models. New information has been developed subsequent to the PVHA elicitation that needs to be evaluated by the DOE for effect on its preferred probability models.

3.2 CONSEQUENCES

The DOE will need to estimate the dose consequences of igneous activity affecting the performance of the proposed repository. Basaltic igneous systems exhibit a wide range of physical characteristics that must be interpreted from sparse, often poorly preserved geologic features in the YMR. In addition, the interactions of basaltic magma with the geologic repository system have no known analog. Dose calculations will require significant extrapolation of igneous process models to the disturbed geologic setting of the repository and to potential interactions with the engineered barrier systems. Staff will review DOE assumptions and models used to estimate the effects of volcanic eruptions and igneous intrusions for consistency with past igneous activity in the YMR and with processes observed at historically active volcanoes analogous to those in the YMR. Staff also will determine if the dose analyses have been performed in a way such that the effects of igneous activity have not been underestimated. The following sections provide information on data and models used to evaluate the consequences of igneous activity in the YMR.

3.2.1 Characteristics of YMR Basaltic Igneous Activity

3.2.1.1 Technical Basis

This section outlines staff's current understanding of the range of physical processes represented by the basaltic igneous systems in the YMR. Because most of these basaltic systems are poorly preserved or exposed in the YMR, igneous processes important to performance must be interpreted from sparse data. Within these limitations, however, staff conclude that the character of past YMR igneous activity represents the most conservative bounds on future YMR activity. In order to test performance models for consistency with past YMR basaltic igneous activity, staff must develop an independent technical evaluation of the range of important processes represented by existing YMR basaltic igneous systems.

Basaltic igneous activity in the YMR since around 8 Ma has encompassed a wide range of processes that effect different implications for repository performance. Many of these processes are interpreted from a sparse, poorly preserved geologic record, especially for basaltic centers older than about 4 Ma. Observations at some older YMR centers, in addition to historically active basaltic volcanoes, indicate that low-energy, low-dispersivity eruptions have limited potential to disperse HLW to critical group locations. Such volcanoes commonly are referred to as hawaiian or low-energy strombolian style and are characterized by small volumes of subsurface disruption, low eruption velocities, and limited dispersal of tephra (e.g., Walker, 1993). The youngest YMR volcanoes and many analogous historically active cinder cones, however, clearly had relatively high-energy, high-dispersivity eruptions with the potential to disperse HLW to proposed critical group locations (Connor, 1993; Hill and Connor, 1995; Hill et al., 1995; Hill, 1996). These eruptions are commonly referred to as violent strombolian style and are characterized by relatively large volumes of subsurface disruption, high eruption velocities, and extensive dispersal of tephra (e.g., Blackburn, et al., 1976; Walker, 1993). Acceptable consequence models will examine in detail the characteristics of violent strombolian basaltic volcanoes, as these eruption styles present the greatest potential hazard to inhabitants located tens of kilometers away from the proposed site.

3.2.1.1.1 Subsurface Conduit Diameters

The diameter of the volcanic conduit controls the amount of HLW available for transport. Conduits for <5 Ma YMR volcanoes are only exposed to depths of several dekameters, which will not accurately represent conduit diameters at 300-m depths. Conduit diameters can be estimated, however, through the volume of shallow wall-rock xenoliths erupted. Xenoliths <0.7 mm in diameter at Lathrop Wells volcano average around 1 volume percent for nonhydromagmatic facies (Crowe, et al., 1986). Staff recently evaluated millimeter-to-decimeter diameter xenolith abundances at Lathrop Wells volcano using image analysis methods. For 17 exposures, each encompassing about 1 m^2, millimeter-to-decimeter diameter xenoliths at Lathrop Wells average 0.9 ± 0.6 volume percent (Doubik and Hill, 1999). Most of these xenoliths are derived from Miocene tuffs, which have an estimated thickness of around 500 m beneath Lathrop Wells volcano (Swadley and Carr, 1987). The Lathrop Wells volcano also is characterized by relatively fragmented cone scoria and lacks significant agglutinate beds, indicating a relatively high-energy eruption (e.g., Hill, 1996). Historically active basaltic volcanoes with cone and tephra-fall characteristics similar to Lathrop Wells have tephra-fall deposits roughly twice the volume of the cone (Segerstrom, 1950; Booth, et al., 1978; Budnikov, et al., 1983; Amos, 1986; Hill, et al., 1998). By analogy, tephra-fall deposits at Lathrop Wells volcano were likely twice the cone volume. Lathrop Wells volcano, thus, produced around 7.2×10^7 m^3 of tephra (Table 3), of which 1 percent was likely composed of tuffaceous xenoliths. Assuming the conduit was cylindrical and the xenoliths were derived from $\leq$ 500 m, this volume corresponds to a 40-m diameter conduit beneath Lathrop Wells volcano. In comparison, 1975 Tolbachik Cone 1 produced a 49 ± 7-m diameter conduit during late-stage disruption (Hill, 1996; Doubik and Hill, 1999). For TSPA-VA analyses, DOE assumed a mean conduit diameter of 50 m, with a log-normal distribution to a maximum conduit diameter of 120 m. Although the mean diameter remains constant, the maximum diameter increases to 150 m in current DOE models (CRWMS M&O, 2000c).

3.2.1.1.2 Eruption Style and Volumes

Several preserved features at Lathrop Wells and Little Black Peak volcanoes indicate a violent strombolian eruption style. First, these volcanoes have unusually high subsurface rock-fragment abundances relative to other Quaternary YMR volcanoes and other basaltic volcanoes in the western Basin and Range. Rock fragments <1 mm average around 1 volume percent at Lathrop Wells (Crowe, et al., 1986). As explained in Section 3.2.1.1.1, millimeter-to-decimeter diameter xenoliths at Lathrop Wells average 0.9 volume percent. Larger rock fragments also appear to be about 0.5 percent at Little Black Peak. In contrast, other typical Basin and Range basaltic volcanoes have less than 0.01 volume percent rock fragments (e.g., Valentine and Groves, 1996). Second, juvenile cone scoria at Lathrop Wells and, to a lesser extent, Little Black Peak consists of angular, broken pieces of larger fragments that were cool on impact with the cone slope. Typically, cinder cone eruptions do not eject material high enough to cool sufficiently to permit brittle fragmentation (e.g., Walker and Croasdale, 1972) whereas violent strombolian eruptions do. Finally, a common strombolian cinder cone feature is beds of agglutinated tephra that accumulated at temperatures high enough to deform plastically and form highly cohesive beds (e.g., Walker, 1993). Lathrop Wells and, to a lesser extent, Little Black Peak consist of loose, nonagglutinated tephra, indicating that these eruptions were more explosive than typical strombolian basaltic volcanoes. Relative to other Quaternary YMR volcanoes, Hidden Cone and the Little Cones also show scoria fragmentation and agglutination characteristics representative of periodically sustained eruption columns and may have had periods of violent strombolian activity.

Although Lathrop Wells and Little Black Peak are the best preserved YMR basaltic volcanoes, remnants of the latest, most potentially disruptive stage of these eruptions are only preserved on the cone flanks. Erosion has removed the upper several meters of the Lathrop Wells tephra-fall deposits (e.g., Crowe, et al., 1995), whereas fall deposits have been completely eroded at Little Black Peak. As documented in Hill (1996) and Doubik and Hill (1999), xenolith breccias indicated that late-stage disruption events likely occurred at Lathrop Wells volcano and possibly at Little Black Peak, analogous to those that occurred during the 1975 Tolbachik eruption (Budnikov, et al., 1983; Doubik, 1997). Because YMR volcanoes older than 1 Ma are extensively eroded, deposits representative of more energetic phases of an eruption may not be preserved. An additional complication in interpreting eruption style from sparsely preserved cinder cone deposits arises from the 1975 Tolbachik eruption. The Cone 2 phase of Tolbachik activity sustained tephra columns 2–4 km high, yet resulted in a highly agglutinated cone that was breeched by a large-volume lava flow (Fedotov, et al., 1984). Although similar degrees of cone agglutination and lava flow volumes are preserved at 1 Ma Crater Flat volcanoes, these features might be consistent with a range of eruption styles that include violent strombolian. Staff conclude that an assumption of a violent strombolian eruption style is reasonable for YMR basaltic volcanoes, and that this assumption will not underestimate risk.

Only sparse and incomplete exposures of tephra-fall remain for Lathrop Wells volcano, which is the youngest and best-preserved YMR volcano (Hill, et al., 1995). With the exception of eroded tephra-fall remnants that occasionally crop out beneath Pliocene lavas and in fault trenches located in and around Crater Flat, tephra-fall deposits have been eroded from other Miocene and younger YMR volcanoes. Tephra-fall volumes for Quaternary YMR volcanoes, however, can be estimated by comparison with fall:cone and cone:lava volume-ratios for well-preserved young basaltic volcanoes. These data are summarized in Table 3. Note that these data are in bulk volume and have not been corrected to dense rock equivalents. Violent strombolian volcanoes have tephra-fall deposit volumes roughly twice that of the volcanic cone, whereas less energetic strombolian cones have roughly equivalent tephra-fall and cone volumes. These relationships are used to estimate fall volumes for Quaternary YMR volcanoes (Table 3). Note that cone:lava ratios for YMR Quaternary volcanoes also encompass the same range as historically active analog volcanoes (Table 3). Using an estimated DRE tephra-fall volume of 2.2×10^7 m^3 for Lathrop Wells, an average mass-flow rate of 25 m^3 s^{-1} (Table 4), and the relationships in Wilson, et al. (1978) and Walker, et al. (1984), the main tephra-producing phase of the Lathrop Wells eruption would have lasted roughly 10 days and produced a column 3.8-km high.

3.2.1.1.3 Magma Fragmentation

In the TSPA-VA, DOE uses a volcanic disruption model based on the depth at which an ascending magma becomes fragmented (i.e., discontinuous particles of magma in a gaseous matrix). CRWMS M&O (1998a) assumes this fragmentation depth is between 100–400 m, based on interpretations of magmatic volatile contents. There are several difficulties with this modeling approach. First, wall-rock xenoliths from depths >400 m are observed at basaltic volcanoes with relatively low degrees of magma fragmentation and tephra dispersivity (Valentine and Groves, 1996). These xenoliths demonstrate that ascending magma can break, entrain, and erupt wall-rocks through mechanisms besides conduit abrasion (e.g., Macedonio, et al., 1994; Doubik and Hill, 1999). As wall-rock behavior is used as a general analog for waste-package behavior (CRWMS M&O, 1998a), similar process below the fragmentation depth thus appear capable of breaking, entraining, and erupting HLW. Second, CRWMS M&O (1998a) concludes that basalt above the fragmentation depth has cooled to ambient temperatures and that only solid particles impact affected waste packages. However, many observed basaltic eruptions have sustained tephra columns supported by a core of incandescent (i.e., temperature >700 °C), fragmented magma. In addition, CRWMS M&O (1998a) does not describe a physical mechanism to rapidly cool large volumes of roughly 1100 °C magma under a two-phase flow regime (e.g., Vergniolle and Jaupart, 1986), where the only significant heat-loss mechanisms are conductive cooling along <400 m of conduit walls and differential flow of the low heat-capacity magmatic gas. Alternatively, the available information indicates little, if any, cooling occurs during the transition to a fragmented magma at depths <100 m. Finally, CRWMS M&O (1998a) uses magmatic water contents as low as 1 weight percent to effect fragmentation depths around 100 m. Also, it should be noted that available experimental data on basalts similar to YMR basalts (Knutson and Green, 1975) clearly demonstrates that YMR magmatic water contents must have been greater than about 2 weight percent to result in observed mineralogical features (e.g., Vaniman, et al., 1982; CRWMS M&O, 1998a). The current DOE modeling approach does not attempt to use magmatic volatile contents to calculate fragmentation depths and instead assumes conditions representative of violent strombolian eruptions occur at repository depths (CRWMS M&O, 2000c, d). Thus, all eruptions in DOE TSPA models are capable of entraining and transporting HLW from disrupted waste packages. Although magmatic volatile contents are not used in the current NRC TSPA modeling approach, these volatile contents can be related to the dispersal capability of basaltic volcanoes with violent strombolian eruption styles (e.g., Roggensack, et al., 1997).

3.2.1.2 Summary

The physical volcanology of YMR basaltic volcanoes is varied but indicates that violent strombolian activity was common and appears characteristic of the most recent eruptions. Violent strombolian eruptions appear capable of widening subsurface conduits to tens of meters in diameter, entraining and dispersing large volumes of wall rock, and transporting tephra at least tens of kilometers down wind. Thus, models of volcanic eruption through the proposed repository need to encompass dose-estimates resulting from this style of volcanic activity. In the TSPA-VA, DOE uses a model based on fragmentation depth that does not appear consistent with violent strombolian activity (CRWMS M&O, 1998a). The current DOE modeling approach does not rely on fragmentation depth and assumes all eruptions are consistent with violent strombolian activity (CRWMS M&O, 2000d).

3.2.2 Tephra Dispersal Models

3.2.2.1 Technical Basis

Acceptable estimates of radiological dose and risk associated with volcanic eruptions through the Yucca Mountain repository depend on numerical models of HLW transport upward in a volcanic tephra column, advection and dispersion of HLW with volcanic ash in the atmosphere, and deposition of HLW in the tephra deposit at a critical group location. The accuracy of these estimates depends on capturing fundamental details of volcanic ash-plume dynamics (e.g., Sparks, 1986; Sparks, et al., 1997), of which there are numerous historical examples from basaltic cinder cone eruptions (Figure 29). Models of volcanic tephra eruptions range from simplistic models that can capture the general pattern of tephra dispersion without attempting to portray the physics of volcanic columns accurately (e.g., Suzuki, 1983), to thermo- fluid-dynamic models of eruption columns and particle advection and dispersion (e.g., Woods and Bursik, 1991; Sparks, et al., 1992, 1997; Woods, 1993, 1995). These latter models make a convincing case that accurate, quantitative descriptions of tephra deposition at the ground surface result from application of physically accurate models. Thus, although computationally complex, these models can likely provide insight into the behavior of HLW in the eruption column despite the very different physical properties of HLW relative to basaltic tephra.

These same arguments for physical detail extend to the sedimentation of tephra and HLW out of the atmosphere. For example, Bonadonna, et al. (1998) have shown that particle Reynolds number plays a critical role in particle settling velocity and, as a result, the particle-size density distributions in the resulting tephra deposit. One of the first attempts to quantify the dispersion of tephra in volcanic eruptions was by Suzuki (1983). Suzuki's model has been modified and applied to volcanic eruptions by Glaze and Self (1991) and Hill, et al. (1998) and applied to the transport of HLW during volcanic eruptions by Jarzemba (1997). In the Suzuki model, the erupting column is treated as a line source reaching some maximum height governed by the energy and mass flow of the eruption. A linear decrease in the upward velocity of particles is assumed, resulting in segregation of tephra or tephra and waste particles in the ascending column by settling velocity, which is a function of particle size, shape, and density. Particles are removed from the column based on their settling velocity, the upward decrease in velocity of the column as a function of height, and a probability density function that attempts to capture some of the natural variations in the parameters governing particle diffusion out of the column. Dispersion of the ash diffused out of the column is modeled for a uniform wind-field and is governed by the diffusion-advection equation with vertical settling.

3.2.2.1.1 Alternative Eruption Column Models

The Suzuki (1983) model does not attempt to quantify the thermo- fluid-dynamics of volcanic eruptions. The more recent class of models, pioneered by Woods (1988), concentrates on the bulk thermophysical properties of the column, defining a gas thrust region near the vent and a convective region above, within which the thermal contrast between the atmosphere and the rising column results in the entrainment of air and buoyancy forces loft particles upward. In contrast to Suzuki (1983), this class of models results in a highly nonlinear velocity profile within the ascending column. This difference can have a profound effect on the ascent height of HLW particles in an ascending eruption column and ensuing dispersion in the accessible environment. Woods (1988; 1995) developed the following method of modeling the physical state for the eruption column. Vertical flux of material in the rising column is given by $\pi L^2 u\beta$, where u, L, and β are column velocity, column radius, and column bulk density, respectively. Air is entrained in the column based on an entrainment coefficient, ε (typically equal to 0.1), and the surface area of the column. In the steady state, conservation of mass in the gas-thrust region of the eruption column is given by

$$\frac{d(uL^2\beta)}{dz} = \frac{uL}{8}\sqrt{\alpha\beta} \qquad (40)$$

where α is the ambient air density and z is vertical distance above the vent. In the convective region of the eruption column, conservation of mass is expressed as

$$\frac{d(uL^2\beta)}{dz} = 2\varepsilon\alpha uL \qquad (41)$$

This formulation does not account for the loss of large particles from the plume that have settling velocities greater than the upward velocity of the plume or are ejected as projectiles from the margins of the column. Woods (1988, 1995) casts the conservation of momentum equation for buoyantly rising volcanic columns as

$$\frac{d(u^2L^2\beta)}{dz} = (\alpha - \beta)gL^2 \qquad (42)$$

where g is gravitational acceleration, and conservation of energy as

$$\frac{d}{dz}(C_p\theta\beta uL^2) = C_aT\frac{d}{dz}(\beta uL^2) + \frac{u^2}{2}\frac{d}{dz}(\beta uL^2) - \alpha uL^2 g \qquad (43)$$

where

$$C_p = C_a + (C_{po} - C_a)\frac{(1 - n)}{(1 - n_o)} \qquad (44)$$

T is air temperature, C_p is the bulk specific heat of the gas column (magmatic gas + entrained air + pyroclasts), θ is the temperature of the column, C_a is the specific heat of air, C_{po} is the specific heat of the magma, n is the gas mass-fraction in the column, and n_o is the gas mass-fraction in the column at the vent. Bulk density of the ascending column is

$$\frac{1}{\beta} = (1 - n)\frac{1}{\sigma} + \frac{nR_g\theta}{P} \tag{45}$$

where σ is the pyroclast density, P is atmospheric pressure, and R_g is the molecular weight of the bulk gas in the eruption column multiplied by the gas constant. The gas mass-fraction is in turn given by

$$n = 1 + (n_o - 1)\frac{(L_o^2 u_o \beta_o)}{L^2 u \beta} \tag{46}$$

where L_o, u_o, and β_o are the initial vent radius, velocity, and bulk column density at the vent and

$$R_g = R_a + (R_{go} - R_a)\left(\frac{1 - n}{n}\right)\left(\frac{n_o}{1 - n_o}\right) \tag{47}$$

where R_{go} and R_a are the products of the gas constant and the molecular weight of gas in the eruption column at the vent and air, respectively. These equations can be recast in terms of three variables, here called *M1*, *M2*, and *M3*, and the three coupled differential equations can be solved numerically for a given set of initial conditions. In the gas thrust region

$$M1 = uL^2\beta \tag{48}$$

$$M2 = u^2L^2\beta \tag{49}$$

$$M3 = C_p\theta\beta uL^2 \tag{50}$$

$$\frac{dM1}{dz} = \frac{uL}{8}\sqrt{\alpha\beta} \tag{51}$$

$$\frac{dM2}{dz} = (\alpha - \beta)gL^2 \tag{52}$$

$$\frac{dM3}{dz} = \left(C_aT + \frac{u^2}{2}\right)\frac{dM1}{dz} - \alpha uL^2g \tag{53}$$

where

$$u = \frac{M2}{M1} \quad (54)$$

and

$$\theta = \frac{\left[\frac{M3}{M1}\right]}{C_p} \quad (55)$$

$$n = 1 + (n_o - 1)\frac{(L_o^2 u_o \beta_o)}{M1} \quad (56)$$

In the convective region of the column

$$\frac{dM1}{dz} = 2\varepsilon\alpha u L \quad (57)$$

and the other equations remain unchanged.

As an example using the initial conditions and constants from Table 1, the gas-thrust region extends to approximately 150 m above the vent. At this point, θ = 921 °K, n = 0.31, u = 58.7 m s^{-1}, and L = 75 m. The plume then becomes buoyant above 150 m and rises to a column height of approximately 4.5 km. At about 4 km, the radius of the eruption column begins to increase rapidly to L > 1 km, and the upward velocity of the column begins to decrease rapidly (Figure 30) as the column reaches neutral buoyancy. Thus, these initial conditions and parameter distributions yield a column height appropriate for the sustained column during a violent strombolian eruption (Figure 29).

Total rise time of the plume, (R_T), is calculated as

$$R_T = \int_{vent}^{(\beta < \alpha)} \frac{1}{u_{jet}(z)} dz + \int_{(\beta < \alpha)}^{(u(z) \to 0)} \frac{1}{u_{conv}(z)} dz \quad (58)$$

and for the above example is approximately 185 s. With wind velocities on the order of 9 m s^{-1}, the center of the column will be displaced approximately 1.6 km down wind between the vent and the level of neutral buoyancy. Based on the vertical velocity profile (Figure 30), nearly all of the horizontal displacement will occur in the upper few hundred meters of the ascending column as the vertical velocity approaches the wind velocity.

An important conclusion from this analysis is that velocities in the eruption column remain high until near the top of the column. As particle transport in the eruption column depends on the bulk properties of the column, there is little opportunity for dense HLW particles to fall out of the erupting column, unless they are advected to the column edge during ascent. This is a different result than predicted from Suzuki (1983), who estimates the height at which material diffuses out of the column as a simple function of the particle settling velocity. Hence, the Suzuki (1983) model predicts that dense HLW particles will tend to be "released" from the eruption column at comparatively low altitudes, resulting in comparatively lower dispersion. In contrast, the thermo-fluid-dynamic model tends to transport HLW and HLW-laden particles to higher altitudes, resulting in wider dispersion of this material. The difference between these models will become more pronounced at higher eruption velocities. Furthermore, parameters like bulk density (Eq. 45) of the column can be modified to specifically examine the dispersion of HLW. Differences between these models may significantly affect dose calculated at critical group locations 20 km from the proposed repository site.

Less energetic stages of a cinder cone-forming eruption produce weak plumes that bend over as they rise due to advection by wind (Figure 29). Sparks, et al. (1997) note that these weak plumes can remain highly organized as they are advected downwind. Such plumes can form convection cells or retain a puffy character with little entrainment and mixing with air. Thus, sedimentation out of these plumes may be slower than expected using the diffusion-advection equation. For example, although the 1995 eruption of Cerro Negro (Figure 29) produced a relatively small volume of tephra (3×10^6 m^3) in a column that rose to only 2–2.5 km, ash-fall deposits 20 km downwind were 0.5 cm (Hill, et al., 1998). Eruptions of this magnitude are capable of effecting peak annual total effective dose equivalents (TEDE) on the order of rems for critical groups located 20 km from a repository-penetrating volcanic eruption. Clearly, reasonably conservative consequence analyses will need to evaluate dose from large, convective eruptions that ascend to atmospheric levels of neutral buoyancy as well as smaller eruptions with column ascent limited by prevailing winds.

3.2.2.1.2 Wind Speed Data

Wind speed is a parameter that significantly affects tephra dispersion models for basaltic volcanoes (e.g., Hill, et al., 1998). The column from the next YMR eruption will likely reach altitudes of 2–6 km above ground level, as is observed for most violent-strombolian basaltic eruptions (e.g., Table 4). Although near ground-surface wind data are available for the proposed repository site, low-altitude winds will be affected significantly by surface topographic effects and, thus, have little relevance to modeling dispersal from 2–6-km-high eruption columns (e.g., U.S. Department of Energy, 1997). The nearest available high-altitude wind data are from the Desert Rock airstrip, which is located about 50 km southeast of Yucca Mountain. Based on data in U.S. Department of Energy (1997), average wind speeds at about 2 km above ground level (i.e., 700 mbar) are 6 m s^{-1}. These average wind speeds increase to about 12 m s^{-1} at altitudes of about 4 km above ground level (i.e., 500 mbar). Staff conclude that an average wind speed of 12 m s^{-1} provides a reasonably conservative basis to model aerial tephra dispersal from the proposed repository site.

For TSPA-VA analysis, DOE used wind speeds and directions obtained from near surface stations (CRWMS M&O, 1998a) to reduce the percentage of time a dose could get to the critical group. This approach does not consider several important factors that could result in calculated doses greater or less then those presented by DOE. First, the near surface data is highly controlled by the local topography, and, therefore, is not representative of the directions and speeds at the altitude of the ash plume (i.e., in the 2–6 km range above ground surface). Second, there are no site-specific data, and the nearest available data (from Desert Rock Airstrip) is constrained to 4 km above ground surface (U.S. Department of Energy ,1997). Finally, any tephra-fall deposit will be reworked and redistributed by near-surface winds following initial deposition. As a result, at least some component of the deposit would likely be displaced to the location of the critical group, providing a dose to the critical group even if the original deposit was not located in that area. To account for potential remobilization effects and significant uncertainties in wind speed and direction data, a reasonably conservative approach for PA analysis is to assume the winds depositing the tephra fall deposit are directed south to the critical group. Current DOE analyses direct the eruption plume toward the critical group for all PA models (CRWMS M&O, 2000d). Wind speeds used in these analyses, however, are determined from Yucca Flat measurements at altitudes <4 km above ground level. Additional data are needed by the DOE to determine appropriate wind speeds for future eruptions occurring in the YMR at 2–6-km altitudes above ground level.

3.2.2.2 Summary

Basaltic eruptions that build cinder cones evince dramatic variations in energy, duration, and style. Numerical models that quantify the physics of these eruptions have reached a stage of development that allows exploration of the parameters governing this variation. Thus, many of the nuances of observed eruption columns and their deposits can now be understood in terms of fundamental physical processes (e.g., Sparks, et al., 1997). Such an understanding is critical for volcanic risk assessment related to the proposed Yucca Mountain repository because there are no observations of the behavior of very dense HLW particles in eruption columns. There also is considerable uncertainty in how to simulate the entrainment and dispersal of HLW in these columns. Physically accurate eruption column models provide an opportunity to extend our understanding of tephra plumes to encompass the distribution and deposition of dense HLW particles in tephra deposits. In these circumstances, application of physically accurate models is a fundamental step in stochastic modeling of dose and risk to a critical group. In the TSPA-VA and subsequent models (CRWMS M&O, 2000d), DOE used the tephra dispersal models of Suzuki (1983) as modified by Jarzemba (1997) and Hill, et al. (1998).

3.2.3 Magma-Repository Interactions

3.2.3.1 Technical Basis

This section outlines how repository construction can potentially interact with and modify the characteristics of the volcanic eruption. Construction and the effects of the repository will cause stress redistribution associated with drift free-surface effects and possibly thermal effects on rock strength associated with waste emplacement. These effects, in turn, affect rates of magma injection into repository tunnels following the dike intersection, the temperature, pressure, geochemical conditions prevailing in the repository following dike injection, and the development of volcanic vents and associated tephra dispersal rates.

Basaltic intrusion propagation is largely controlled by the distribution of stress in the shallow (i.e., <10 km) crust (e.g., Delaney, et al., 1986). The emplacement of 5- to 10-m diameter drifts at 300-m depths represents a free surface that will likely affect the distribution of crustal stress for some distance around the drifts. The upward ascent of basaltic magma may be affected by this stress redistribution, resulting in ascent characteristics that are not reasonably analogous with magma ascent in undisturbed geologic settings. Lateral intrusion propagation also may be affected by this stress redistribution, which affects the area disrupted by an igneous event.

In addition to stress redistribution, the repository drifts represent free surfaces where lithostatic confining pressure is zero. Ascending basaltic magma, which contains dissolved volatiles, will be under roughly 10 MPa lithostatic confining pressure when it encounters the drifts. Nonequilibrium decompression will ensue, resulting in rapid volatile exsolution (e.g., Connor and Hill, 1993b). Although the magnitude and consequences of this rapid exsolution have not yet been modeled, volatile expansion and magma fragmentation are often related to conduit erosion and wall-rock entrainment (i.e., Macedonio, et al., 1994; Valentine and Groves, 1996).

The approach used to address this complex problem of magma-repository interactions was to first perform analytical calculations that help bound conditions during and following magma injection. These calculations are not intended to model or predict exact conditions within the repository during igneous events. Rather, these analytic calculations are intended to help identify which processes require further analyses using more sophisticated numerical and experimental techniques. The initial scoping calculations for (i) flow conditions during injection of volatile-poor and volatile-rich magmas into repository drifts, and (ii) fracturing of the rock above the drifts due to pressure conditions following magma injection. Results of these initial calculations are summarized in the following sections. Temperature conditions within repository drifts following magma injection are discussed in Section 3.2.4.1.1.

3.2.3.1.1 Flow Conditions

Woods and Sparks (1998) performed initial scoping calculations for magma flow inside a repository drift. To estimate flow conditions in a repository drift, Woods and Sparks (1998) assumed that a 1–2 m wide dike originates from a magma reservoir at a depth of 5–10 km and intersects the drift. An overpressure within the dike system of 1–10 MPa is required to propagate this dike upward and to maintain a dike fracture-width of 1–2 m. Assuming that the drifts are located at a depth of 300 m, where lithostatic pressure is on the order of 10 MPa, the total pressure at the dike tip just prior to breaking into this drift also is on the order of 10–20 MPa. Pressures in the drift are assumed to be much less than lithostatic, on the order of atmospheric pressures (i.e., 0.01 MPa). Under such conditions, magma would probably be diverted into the horizontal drifts. Sample calculations were performed for volatile-free magmas using typical basaltic magma viscosities of 10–100 Pa s. Under initial pressure conditions of 5–20 MPa and dike widths of 1–2 m, magma is expected to accelerate to 1–20 m s^{-1} as it enters the drift, potentially filling the drift in several tens of seconds. In response to this acceleration at the magma front, however, pressure within the dike is expected to decrease and the dike may narrow or collapse after this initial acceleration of magma within the drift. Under these circumstances, magma injection may become pulsed, with overpressure building at the dike tip, followed by periodic injection of magma into the drift.

Initial scoping calculations also were made for basaltic magmas containing 1–3 weight percent water at high pressures (i.e., Section 3.2.1.1). During decompression events, such as when the magma intersects repository drifts at near atmospheric pressure, the volatiles become supersaturated and exsolve from the magma. This exsolution of volatiles increases the volume of the magma-gas mixture, decreases the mixture’s density, and yields a compressible flow. These changing conditions accelerate flow of the magma-gas mixture into the drift. Initial calculations suggest that a shock wave can develop at 8–9 times atmospheric pressure and propagate through the tunnel at speeds of up to 150 m s^{-1}. This also is a typical flow velocity for normal strombolian-style volcanic eruptions, observed at the earth's surface during the eruption of basaltic magmas similar in composition to Quaternary basalts of the YMR.

These initial scoping calculations do not capture the true complexities of high pressure and temperature flows. Nonetheless, the results of these calculations indicate that basaltic magma will be diverted and accelerate into drifts during dike injection. Ongoing numerical and physical analog experiments should reveal significant details about these flow conditions.

3.2.3.1.2 Fracturing of Drift Walls

If a significant amount of the magma intersecting the repository is redirected into drifts, it becomes necessary to consider conditions following magma injection that may lead to the development of volcanic eruptions at the surface above the repository. In the following calculations, the fluid pressures required to initiate vertical fracturing above the drift and upward propagation of magma are estimated using the Kirsch solution for rock hydrofracturing (e.g., Goodman, 1980).

The Kirsch solution solves for the compressive stresses near a conduit or drift:

$$\sigma_{rr} = \frac{\sigma_1 + \sigma_2}{2}\left(1 - \frac{a^2}{r^2}\right) + \frac{\sigma_1 - \sigma_2}{2}\left(1 - \frac{4a^2}{r^2} + \frac{3a^4}{r^4}\right)\cos 2\theta \qquad (59)$$

$$\tau_{r\theta} = -\frac{\sigma_1 - \sigma_2}{2}\left(1 + \frac{2a^2}{r^2} - \frac{3a^4}{r^4}\right)\sin 2\theta \qquad (60)$$

$$\sigma_{\theta\theta} = \frac{\sigma_1 + \sigma_2}{2}\left(1 + \frac{a^2}{r^2}\right) - \frac{\sigma_1 - \sigma_2}{2}\left(1 + \frac{3a^4}{r^4}\right)\cos 2\theta \qquad (61)$$

where

a is the conduit or drift radius
r is the distance from the center of the conduit ($r \geq a$)
θ varies from 0 deg in the direction of σ_1 and 90 deg in the direction σ_2
σ_1 is the greatest principal compressive stress, in the case of a drift $\sigma_1 = \sigma_{max,\ hort}$
σ_2 is the least principal compressive stress, in the case of a drift $\sigma_2 = \sigma_{min,\ hort}$
σ_{rr} is the radial compressive stress
$\sigma_{\theta\theta}$ is the tangential compressive stress
$\tau_{r\theta}$ is the shear stress

Tangential compressive stress is minimum along the σ_1 axis at the drift wall and is

$$\sigma_{\theta\theta} = 3\sigma_2 - \sigma_1 \tag{62}$$

and the tangential compressive stress is at a maximum along the axis at the drift wall and is

$$\sigma_{\theta\theta} = 3\sigma_1 - \sigma_2 \tag{63}$$

The magnitudes of displacements are also derived from the Kirsch solution assuming elastic behavior and are

$$u_{rr} = \frac{\sigma_1 + \sigma_2}{4G}\frac{a^2}{r} + \frac{\sigma_1 - \sigma_2}{4G}\frac{a^2}{r}\left[4(1-v) - \frac{a^2}{r^2}\right]\cos 2\theta \tag{64}$$

and

$$u_{\theta\theta} = -\frac{\sigma_1 - \sigma_2}{4G}\frac{a^2}{r}\left[2(1-2v) + \frac{a^2}{r^2}\right]\sin 2\theta \tag{65}$$

where

u_{rr} is the radial outward displacement
$u_{\theta\theta}$ is the tangential displacement
G is the shear modulus (i.e., modulus of rigidity)
v is Poisson's ratio

For a typical host rock, Lister and Kerr (1991) use $G = 1 \times 10^{10}$ Pa. Alternatively, Pollard (1987) argues for a much lower value $G = 1 \times 10^9$ Pa and $v = 0.25$. These differences in G may be important for models at repository depths.

Assume that the drift is 1 km long, G is 1×10^9 Pa, and v = 0.25. For a one-meter-wide dike to propagate to the surface, the driving pressure (P-S) = 1.3×10^6 Pa. This result is highly dependent on a value of G, which could be as high as 1×10^{10} Pa, indicating that 1×10^6 Pa < (P-S) < 1×10^7 Pa. This suggests that a fluid pressure of at least 3 MPa is needed to form a 1-m-wide dike along the length of the drift trending perpendicular to the least horizontal compressive stress and at least 5–6 MPa is needed to form a 1-m-wide dike along the length of the drift trending perpendicular to the maximum horizontal compressive stress. These fluid pressures might need to be one order of magnitude higher depending on the value assumed for the shear modulus. Measured values of Young's modulus and Poisson's ratio at Yucca Mountain are 27–32 GPa and 0.21, respectively, giving G as 1–1.3×10^{10} Pa for the intact rock mass (CRWMS M&O, 1997), which indicates higher fluid pressures are required. Thermal conditions around the magma-filled drift may lower the value of G (Pollard, 1987).

If the magma pressure in the drift is p_f, then an additional stress of magnitude p_f is added everywhere around the drift wall. In order for a new tensile fracture to form, the tensile stress along the σ_1 axis (where tangential compressive stress is minimum) must equal the uniaxial tensile strength of the rock, T_o, and

$$p_f = T_o + 3\sigma_2 - \sigma_1 \qquad (66)$$

where p_f is the fluid pressure in the drift and T_o is the tensile strength of the rock. Note that if the rock is already jointed, then the effective tensile strength is reduced by some factor. For a horizontal drift in tuff trending perpendicular to the regional maximum horizontal compressive stress at 300 m depth

$\sigma_1 = \rho gh = 2600 \times 9.8 \times 300 = 7.6 \times 10^6$ Pa
$\sigma_2 = 5.5 \times 10^6$ Pa (in the YMR σ1:σ2:σ3 = 90:65:25.)
$T_o = 1 \times 10^6$ Pa
$p_f = 9.9 \times 10^6$ Pa

to form a vertical fracture along the length of the roof for a drift trending perpendicular to the regional maximum horizontal compressive stress.

For a horizontal drift in tuff trending perpendicular to the regional minimum horizontal compressive stress at 300 m depth:

$\sigma_1 = \rho gh = 2600 \times 9.8 \times 300 = 7.6 \times 10^6$ Pa
$\sigma_2 = 2.1 \times 10^6$ Pa
$T_o = 1 \times 10^6$ Pa
$p_f = -2.5 \times 10^5$ Pa

to form a vertical fracture along the length of the roof for a drift trending perpendicular to the regional minimum horizontal compressive stress. The negative value for p_f in this case indicates that tensile fractures will likely already exist in a drift trending perpendicular to the regional minimum horizontal compressive stress (roughly N25E in the YMR); therefore, the magma will not need excess fluid pressure to form these fractures.

Note that sill formation is less likely because $\sigma_1 \gg \sigma_2$ in Yucca Mountain at repository depths and in order to propagate the horizontal fracture $p_f > \sigma_1 + T_o$. Thus, based on these calculations, a volcanic eruption will likely follow magma injection into the repository. A caveat to this result is that materials lining the drift walls may have significantly different mechanical properties and could impede the development of fractures if the drift walls were intact.

The magma-driving pressures required to open a conduit to the surface along the length of the tunnel can be estimated using techniques developed by Pollard (1987)

$$\frac{t}{l} \approx \frac{P - S}{G/(1-\nu)} \qquad (67)$$

where

t is the dike thickness
l is the dike length (in this case equal to the length of the tunnel)

P-S is the driving pressure
G is the shear modulus
v is Poisson's ratio

Using a drift length of 1 km, $G = 1 \times 10^9$ Pa, and $v = 0.25$. For a 1-m-wide dike to propagate to the surface, the driving pressure (P-S) = 1.3×10^6 Pa. This result is highly dependent on the value of G, which could be as high as 1×10^{10} Pa, indicating that 1×10^6 Pa < (P-S) < 1×10^7 Pa. Variations in G suggest that a fluid pressure of at least 3 MPa is needed to form a 1-m-wide dike along the length of the tunnel trending perpendicular to the least horizontal compressive stress and at least 5–6 MPa is needed to form a 1-m-wide dike along the length of the tunnel trending perpendicular to the maximum horizontal compressive stress. These fluid pressures might need to be one order of magnitude higher depending on the value assumed for the shear modulus (G).

These overpressures are on the same order as those required to initiate dike propagation from great depth and are greater than or comparable to the overpressures required to initiate hydrofracturing of the tunnel roof based on the Kirsch solution. This result suggests that if (i) most or all of the magma in the dike is redirected into drifts and (ii) a flow path into the repository is reestablished after the initial disruption associated with rapid magma acceleration in the drifts and drainage of the dike, then dike injection may be initiated vertically above the drift along its entire length. For Yucca Mountain drifts, vertical hydrofracturing in the drift roof is much more likely to occur in north-trending tunnels, but fluid pressures needed to initiate hydrofracturing in east-trending tunnels may also be reached. Given the comparatively shallow depth of the tunnels, exsolution of volatiles within the tunnel may also increase the fluid pressure available to inject a 1-m-wide dike to the surface.

These calculations suggest that, following disruption of the repository by injection of magma, volcanic eruptions are likely to be initiated at the surface above the repository, assuming that sufficient magma volume is available to drive the eruption. These calculations also suggest that the location of conduits above the repository and volcanic vents at the surface may be controlled by drift geometry, rather than only by dike geometry. Under these circumstances, the character of eruption columns, used in modeling ash and waste dispersion, may be influenced by this geometry of the shallow conduit system.

3.2.3.2 Summary

The repository itself, potentially affects the shallow subsurface ascent of magma. These effects include change in the depth of volatile exsolution, resulting in potential changes in eruption style, and changes in intrusion geometry. As work in these areas is ongoing, staff have only completed initial scoping calculations for flow conditions during injection of volatile-poor and volatile-rich magmas into repository drifts, and pressure conditions required to initiate fracturing of rock above drifts following magma injection. These scoping calculations indicate that basaltic magma will be diverted and accelerate into open repository drifts during dike ascent. In addition, this magma may have sufficient volume and fluid pressure to propagate vertical fractures along the roof of intersected drifts. These fractures could localize volcano formation over the intersected drifts. In the TSPA-VA, DOE did not evaluate the effects of the repository on magma ascent characteristics (U.S. Department of Energy, 1998b). Initial models were developed in CRWMS M&O (2000e) for a high thermal load repository containing backfilled drifts. These models concluded that magma-repository interactions are potentially significant, and more detailed models are needed to evaluate the effects of these interactions.

3.2.4 Interaction of Magma with Waste Packages and Waste Forms

3.2.4.1 Technical Basis

DOE performance assessments prior to the TSPA-VA have all assumed that the waste package fails on contact with basaltic magma (Link, et al., 1982; Barnard, et al., 1992; Barr, et al., 1993; Wilson, et al., 1994). The general physical characteristics of basaltic magma exceed the design criteria commonly applied to HLW emplacement canisters, such that canister failure appears to be a reasonable, though conservative, initial assumption. For example, basaltic magma in the YMR has an initial temperature of around 1100 °C (i.e., Vaniman, et al., 1982; Knutson and Green, 1975). Assuming no external stress, such as that induced by magma flow, 2.5Cr-1Mo steel will fail through intergranular creep rupture alone at these temperatures at time scales equivalent to the duration of historical basaltic volcanic eruptions (Fields, et al., 1980; Viswanathan, 1989). Ascending basaltic magma also has a nonvesiculated density around 2600 kg m^{-3} and likely impacts the HLW canister between 1–100 m s^{-1}, creating significant external stress that will enhance failure through ductile fracturing (e.g., Ashby, et al., 1979). In addition, basaltic magmatic oxygen fugacities commonly are 10 log units below atmospheric conditions (e.g., Carmichael and Ghiorso, 1990), which may affect Fe^{+2}/Fe^{+3} and Ni/NiO phase relationships in the canister. In addition, basaltic magmas may contain around 0.1 weight percent sulfur, which is readily degassed from the magma at low pressures (e.g., Carroll and Webster, 1994) and likely will affect nickel and chrome alloy phase relationships. A HLW canister failure thus appears reasonably likely for canisters directly intersected by a volcanic conduit. Canisters in contact with basaltic magma introduced through dikes and intradrift lavas may also fail, although thermal and mechanical loads are much lower than those encountered in the volcanic conduit area.

3.2.4.1.1 Canister Heating by Magma

Assuming magma injection into the repository, waste canisters may fail due to mechanical load, chemical corrosion, and thermal load. In several respects, thermal load on the canisters is the simplest of these adverse conditions to evaluate. Heating of the canister by submersion in magma may result in the failure of the canister. Preliminary calculations by CRWMS M&O (1998a) of the TSPA-VA design suggest that canister failure will occur around 800 °C. The behavior of proposed canister materials at magmatic temperatures (around 1100 °C) is poorly known because high temperature tests have not been performed on proposed canister materials and because final canister design currently is not known. In the following analysis, temperatures in the canister are calculated after the intrusion of basaltic magma into the repository drifts and compared to the rate of magma cooling in a drift. The intent of these calculations is to bound the temperature conditions within the canister following magma injection. These calculations are simplified by (i) assuming that the canisters and drifts are infinite in length, (ii) using bulk thermal conductivities for the canisters and wall rock, rather than attempting to account for heterogeneties in these materials, and (iii) assuming canisters are completely submerged in a convecting magma within the drift.

For simplicity, it is assumed that the canister is instantaneously submerged in the convecting magma. Heat transfer within the canister will follow the equation (Carslaw, 1921):

$$\frac{\partial T}{\partial t} = \frac{\alpha}{r}\frac{\partial}{\partial r}\left(r\frac{\partial T}{\partial r}\right) \quad (68)$$

where T is temperature, t is time, α is the thermal diffusivity, and r is radius, with initial conditions

$$T(r, t=0) = T_i \quad (69)$$

and a convective boundary condition at the surface of the canister, at $r = r_o$

$$-k\frac{\partial T}{\partial r} = h_o(T - T_o) \quad (70)$$

White (1984), following Schneider (1955), gives the solution to this heat transfer equation as

$$\theta_c = \frac{T - T_o}{T_i - T_o} = C_1 e^{-\beta_1^2 \alpha t / r_o^2} \quad (71)$$

along the centerline of the cylinder and

$$\theta = \theta_c J_o(\beta_1 r / r_o) \quad (72)$$

off the centerline of the cylinder, for all times greater than

$$t > 0.2 r_o^2 / \alpha \quad (73)$$

In this formulation, J_o is a Bessel function of the first kind. Eq. (71) may also be written using Bessel functions, but here is written using Heisler coefficients, C_1 and β_1 for the centerline formula. These coefficients depend on the Biot number (*Bi)*:

$$Bi = h_o r_o / k \quad (74)$$

and are tabulated in White (1984). Approximation of heat transfer in this way only results in errors at short times [Eq. (73)] after immersion. Thermophysical properties of the canister, basaltic magma, and wall rock are taken from Manteufel (1997) and McBirney (1984) (Table 2).

Temperature of the canister as a function of radial distance and calculated at 30 min intervals is shown in Figure 31. The initial canister temperature is assumed to be 250 °C, and the magma temperature is 1100 °C and does not change for the duration of the calculation (i.e., magma is an infinite heat reservoir). For this calculation, it is assumed that high Biot numbers persist for the duration of heating (Bi = 50, β_1 = 2.35, and C_1 = 1.65). This implies a high heat transfer coefficient between the magma in the tunnel and the canister wall. The heat transfer coefficient, however, may be strongly reduced by formation of a chilled basalt rind on the outer canister wall. Using a low Biot number (Bi = 0.1, β_1 = 0.44, and C_1 = 1.02) results in much slower heating of the canister and produces much lower temperature gradients inside the canister because the heat flux into the canister is limited by h_o (Figure 32).

The same equations can be used to calculate rate of cooling of the magma inside the tunnel, neglecting latent heat of crystallization, the heating of the wall rock, or flow in the tunnel. Here, the heat transfer coefficient is given by

$$h_o = \frac{3.5k}{D} \tag{75}$$

where D is the hydraulic diameter of the tunnel. This gives Bi = 1.75 (β_1 = 1.5 and C_1 = 1.3). For the 5-m-diameter tunnel, temperatures remain near 1100 °C for more than 20 days (Figure 33), indicating there is ample time to heat the canisters to the point of failure during a typical basaltic eruption.

In TSPA-VA, DOE concludes that the waste package will not fail significantly during a volcanic eruption until the inner corrosion resistant material has degraded to <50 percent of original thickness (CRWMS M&O, 1998a). Waste package failure mechanisms evaluated in CRWMS M&O (1998a) are corrosion by volcanic gases, mechanical collapse, and internal pressurization. These analyses used corrosion rates 10^4 greater than used in basecase TSPA-VA analyses, based on 800 °C data in Wang and Douglass (1983). Mechanical collapse was modeled by extrapolating critical-stress temperature dependencies for Alloy 625 from a data range of 20–430 °C to a presumed magmatic temperature of 1000 °C (CRWMS M&O, 1998a). Neglecting all external stress imparted by dense (2600 kg m^{-3}) magma impacting the canister at velocities 10–100 m s^{-1}, CRWMS M&O (1998a) concludes a waste package must be at <50 percent of original thickness to fail through mechanical collapse. Staff note that this analysis has not considered the considerable dynamic stress on the waste package induced by the dense, flowing magma within the volcanic conduit, and that alloy behavior at temperatures ≤430 °C cannot be readily extrapolated to temperatures >1000°C (e.g., Ashby, et al., 1979). Finally, although CRWMS M&O (1998a) concludes that waste package end-cap failure is likely at temperatures >800 °C, staff notes that DOE appears to assume that HLW apparently cannot be entrained from a waste package with intact container walls in subsequent mechanical models. In contrast to the analysis for direct volcanic disruption, CRWMS M&O (1998a) concludes that exposure to magmatic temperatures of 870 °C for 100 hr results in waste package failure for the enhanced source-term scenario. Staff agree that this conclusion appears reasonable for a temperature of 870 °C and note that temperatures more representative of magmatic temperatures (i.e., around 1100 °C) would increase waste package failure in this model. Current DOE models assume waste packages intersected directly by a volcanic conduit will make all the contained HLW available for transport under violent strombolian eruption conditions (CRWMS M&O, 2000d; 2000f). For igneous intrusions intersecting a backfilled drift, DOE concludes magma will flow up to 15 m from the point of intersection (CRWMS M&O, 2000f). All packages impacted by magma within that zone are assumed to loose containment and allow subsequent HLW remobilization through aqueous transport (CRWMS M&O, 2000d).

3.2.4.1.2 HLW Particle Size

In addition to affecting the emplacement canister, the physical conditions associated with ascending basaltic magma will likely affect HLW form. This is important because particle size will directly affect how HLW is incorporated and dispersed during a volcanic event. Particle size also will determine the dosimetry effected through inhalation of contaminated tephra and discrete HLW particles. The high temperature, reducing conditions associated with basaltic magma will likely result in a reduction in spent fuel particle-size through fracturing along grain boundaries and transgranular fracturing (e.g., Ayer, et al., 1988; Einzinger, 1994; Einzinger and Buchanen, 1988). As magma fragments during ascent, particle size will be decreased further through shear induced by conduit flow and volatile expansion. Cooling and atmospheric mixing will occur rapidly in the column (e.g., Thomas and Sparks, 1992), inducing additional thermal and chemical stress on the waste particles. These rapid and relatively large changes in temperature and oxygen fugacity also will likely affect the oxidation state of the HLW, which can affect the mobility of actinide elements at surficial conditions. Process models that calculate the dose consequences of igneous activity will need to account for how the physical conditions of a volcanic eruption affect HLW form.

During the 1960s, the U.S. government developed nuclear-power rocket engines that operated at temperatures comparable to basaltic igneous events (500–1500 °C). Literature from this program was reviewed to determine if there was reasonable analogy with potential HLW behavior during igneous events. The nuclear rocket engines used a reactor core consisting of hollow hexagonal tubes made from 1–7 percent UO_2-Y_2O_3-ZrO_2 fuel in a ceramicized BeO matrix (Cahoon, et al., 1962). Although these tubes were stable at pressures of 342 psi and temperatures of 1454 °C (Lorence, 1973), they do not appear chemically or mechanically analogous to HLW potentially exposed to basaltic magma. Spent reactor-fuel pellets consist of 100 percent UO_2 and associated fission products and are formed from pressed powders having initial particle sizes around 1 μm. They lack a BeO matrix and are not ceramicized, both of which will enhance high-temperature stability significantly. Behavior of nuclear-rocket fuel during engine operation, thus, does not appear reasonably analogous to behavior of HLW during igneous disruptive events.

In the TSPA-VA (CRWMS M&O, 1998a), DOE used *in situ* HLW particle-size distributions from Jarzemba and LaPlante (1996). These particle-size distributions were used in a preliminary analysis for volcanic disruption and did not consider particle-size degradation induced by mechanical, thermal, or chemical processes during igneous events, as outlined above (e.g., U.S. Nuclear Regulatory Commission, 1998b). These effects are important because the TSPA-VA uses a kinetic energy transfer model to entrain HLW from a breeched waste package. In this model, 50 percent of the simulations in the TSPA-VA had HLW particle sizes that were too large to entrain from a breeched container. Of the remaining 50 percent that entrained HLW, 70 percent of the simulations had eruption velocities that were too low to eject HLW from the volcano (CRWMS M&O, 1998a). The use of larger than expected HLW particle sizes in TSPA-VA likely significantly underestimates the amount of HLW potentially dispersed during a volcanic eruption. Subsequently, DOE considered the impact of physical conditions representative of YMR igneous events on HLW particle size (CRWMS M&O, 2000g). Although data are limited, DOE concluded that unaltered spent nuclear fuel will disaggregate during igneous events to average particle diameters of around 20 μm, with a range in particle size of 1–50 μm (CRWMS M&O, 2000g). Staff will conduct sensitivity studies to determine if changes in average particle diameter from 10 μm to 20 μm are significant. These differences in particle diameter, however, do not appear significant based on the relatively large size of tephra particles used to incorporate and disperse HLW for volcanic events.

3.2.4.2 Summary

Available information suggests that the waste package will not be an effective deterrent to the transport and dispersion of HLW during volcanic eruptions. Additional analyses of waste package behavior at high temperature and high mechanical loads may provide new insights, however, a reasonably conservative interpretation of available data is that the waste package fails during a volcanic eruption. Volcanic disruption analyses in TSPA-VA did not consider physical conditions representative of YMR basaltic volcanic eruptions and used lower-temperature data from analog waste package materials to conclude waste package resiliency when exposed to an erupting volcanic conduit. In contrast, TSPA-VA analyses for the enhanced source-term scenario conclude exposure to an intradrift lava flow imparts a thermal load sufficient for waste package wall failure. Staff analyses support this conclusion. The TSPA-VA analyses, thus, do not provide reasonable technical basis that a waste package remains intact when exposed in an erupting volcanic conduit. Staff conclude that waste package failure during igneous events remains a reasonably conservative interpretation of available information. In contrast to the TSPA-VA, current DOE models conclude all HLW is available for transport from waste packages intersected directly by a volcanic conduit or directly impacted by magma flowing from a dike that intersects a repository drift.

Current analyses also suggests that HLW particle fragmentation will occur during a volcanic eruption, due to the mechanical, thermal, and chemical loads imparted on HLW during igneous events. These processes will likely reduce the average HLW particle size significantly below that observed in undisturbed HLW forms. HLW entrainment and transport models in the TSPA-VA did not consider these processes and thus likely underestimated the amount of HLW transported into the accessible environment during volcanic events. In contrast to the TSPA-VA, current DOE models conclude significant reductions in HLW particle diameters occur during basaltic igneous events.

3.2.5 Post-Eruption Processes

3.2.5.1 Technical Basis

Following eruption of the tephra-fall deposit, most of the radiological dose will be acquired through the inhalation of HLW-contaminated ash particles. Few data are available in the literature to evaluate airborne particle concentrations likely to occur above undisturbed basaltic tephra-fall deposits or how these concentrations may change when disturbed by the farming habits characteristic of the critical group. The proposed critical group location of 20 km south of the proposed repository site also is in an area that can be affected by erosion and deposition processes from the Fortymile Wash drainage system. In addition to water, wind can also remobilize contaminated ash particles from other areas and deposit the particles at the proposed critical group location. During long periods of time, remobilization by wind and water, in addition to changes in deposit character through physical and chemical processes, can significantly affect airborne particle concentrations and resulting radiological dose to the critical group. The following sections describe ongoing work in quantifying posteruption modifications to the contaminated tephra-fall deposit.

3.2.5.1.1 Airborne Particle Concentrations

Individuals located 20 km downwind from a repository-penetrating volcanic eruption would receive a radiological dose primarily through inhalation of contaminated ash particles. Particles <200 μm in diameter are resuspended through wind-shear, saltation, and mechanical disturbance of the deposit (e.g., Watson, 1989). A mass-loading model describes the amount of contaminated ash in airborne suspension and is controlled by two critical parameters: airborne mass load and thickness of the surficial deposit capable of eolian entrainment.

Mass load is defined as the airborne mass of particulates per unit volume of air and consists of two primary components (i) airborne mass composed of particles less than 10 μm in diameter, which can be inhaled directly into the pulmonary regions of the lung (i.e., respirable fraction), and (ii) airborne mass composed of particles 10–200 μm in diameter, which are deposited in the naso-pharynx and tracheal-bronchial regions of the respiratory tract upon inhalation. Airborne particle concentrations can range from 10^{-7} to 10^{-4} g m^{-3} for tropical to temperate climates (e.g., Tegen and Fung, 1994) and from 10^{-6} to 10^{-1} g m^{-3} for more arid climates (e.g., Sehmel, 1977; Anspaugh, et al.,1975). Internal dosimetry of the inhaled particles depends on depositional site within the respiratory tract. Studies of nonbasaltic eruptions indicate that inhalation of fine-grained particles may represent a significant health risk (e.g., Baxter, et al., 1999). This suggests that basaltic eruptions could result in a relatively large opportunity for inhalation doses.

Airborne particle concentrations available in the literature are derived from geological deposits that have limited applicability to basaltic tephra-fall deposits. In addition, little information is presented in most of the relevant literature to discern particle size distributions for suspended and surficial deposits, degree of soil development or soil type, vegetative cover, wind conditions, or soil moisture content. This information is necessary to address the suitability of published airborne particle concentrations in evaluating inhalation dose for volcanic deposits. Based on general soil characteristics from the studied environments, however, these soils likely contain significantly lower abundances of suspendable fine particulates than occur in basaltic volcanic fall deposits. These nonvolcanic deposits appear depleted in suspendable fine-grained particulates, represent evolved soil types, and occur in significantly vegetated areas. Based on these characteristics, airborne particle concentrations for these deposits may significantly underestimate the amount of suspendable fine particulates, and, thus, the inhalation dose associated with basaltic volcanic fall deposits.

In some arid environments, some nonvegetated soils and deposits have general grain-size characteristics that might be compared with the volcanic fall deposits. Dune sands, for example, commonly have average grain-sizes comparable to distal volcanic falls (i.e., 150–300 μm); however, the amount of particles <60 μm is often <1 weight percent (e.g., Watson, 1989), much lower then expected from basaltic fall deposits.

Some data are available on airborne particle concentrations following silicic volcanic eruptions. After the 1980 eruption of Mount St. Helens, U.S.A., airborne concentrations of 2–15 μm particles measured about 400 km east of Mount St. Helens ranged from 10^{-2} g/m^3 for several days after the eruption to about 10^{-5} g/m^3 following a significant amount of rain fall (Gage, et al., 1982). Deposit characteristics, height of measurements, and amount of surface disturbing activities were not discussed by Gage, et al. (1982). After the 1995 eruption of Montserrat volcano in the British West Indies, Baxter, et al. (1999) measured average airborne concentrations for <10 μm particles that commonly exceeded >5 × 10^{-5} g/m^3. Human activity resulted in significantly higher airborne particle concentrations than occurred over undisturbed deposits, although actual concentrations were not reported.

To better understand the characteristics of basaltic fall deposits, fresh basaltic volcanic fall deposits were collected 21 km from the vent during the 1995 Cerro Negro, Nicaragua eruption. Preliminary analysis of the Cerro Negro fall deposits indicates that about 2 weight percent of the deposit consists of particles less than 10 μm in diameter, with particles <60 μm constituting about 10 weight percent of the deposit and those <200 μm constituting 50 weight percent of the deposit. Other fall deposits from larger basaltic cinder cone eruptions may contain 2–5 weight percent with diameters <10 μm at 20 km distances (Segerstrom, 1950; Budnikov, et al., 1983; Amos, 1986). Basaltic volcanoes may also produce unusually fine-grained deposits late in the eruption during subsurface brecciation events (Hill, 1996). These types of deposits from the 1975 Tolbachik eruption have more than 40 percent of the associated particles smaller than 60 μm (Doubik, 1997). Similar late-stage, conduit-widening events likely occurred at the youngest YMR volcanoes (Hill, 1996). The largest amount of HLW entrainment would probably occur during this type of event, when the subsurface conduit expanded to dekameters in diameter. Thus, a reasonably conservative risk assessment needs to consider the airborne particle concentrations associated with tephra-fall deposits arising from these conduit widening events, in addition to normal violent-strombolian tephra-fall deposits.

Airborne particle concentrations were measured 1.5 m above basaltic tephra-fall deposits from Cerro Negro volcano, Nicaragua. Concentrations were measured in February 1999 on tephra deposits erupted in early December 1995. More than 2 m of rain fell on these deposits in October 1998 as a result of Hurricane Mitch. Particle concentrations were measured with multi-stage virtual impactors for 4, 10, and 100 μm diameter particles and with high-precision filters for total suspendable particulates. About 60 percent of the airborne particles were 10–100 μm in diameter. Wind speeds measured 1.5 m above ground level averaged 4±2 m/s during the measurements. Deposits that were undisturbed by surface activity had average airborne total particle concentrations on the order of 10^{-4} g m^{-3}. Deposits disturbed by light activity such as walking had average airborne total particle concentrations on the order of 10^{-3} g m^{-3}. Average airborne total particle concentrations on the order of 10^{-2} g m^{-3} were measured while driving over the tephra deposits in an open truck. Detailed results of this investigation are being prepared for publication in FY2001.

Using data from the most reasonably analogous deposits in the available literature (Anspaugh, et al., 1975; Tegen and Fung, 1994), and comparing these data to the previous information on basaltic fall deposits, the staff have determined that airborne particle concentrations of 10^{-4} to 10^{-2} g m^{-3} can be used to describe the initial amount of resuspended particles above a fresh basaltic tephra fall.

In the TSPA-VA, the DOE did not use airborne particle concentrations specific to a basaltic fall deposit (CRWMS M&O, 1998b). Although airborne particle concentrations are not identified for the tephra-fall deposits, CRWMS M&O (1998b) used an average airborne particle concentration of 1.9 × 10^{-5} g m^{-3} for other dust inhalation scenarios. The airborne particle concentrations used in dose modeling in the TSPA-VA appear to significantly underestimate the amount of inhalable and respirable particulates suspended over undisturbed and mechanically disturbed tephra deposits. Current DOE models, however, use an average airborne particle concentration of 10^{-3} g m^{-3} for 10.75 hours per day (CRWMS M&O 2000h). NRC and DOE each are developing a technical basis to model how initial airborne particle concentrations may change with time due to changes in particle-size characteristics of the tephra deposit.

3.2.5.1.2 Fall-Deposit Evolution

Fall-deposit characteristics will change with time as the deposit is exposed to subaerial environmental conditions. The amount of resuspendable ash particles likely will decrease through time by wind elutriation and rainwater infiltration. In addition, the fall deposit will be eroded through sheet-wash and channelized surficial flow. Erosion, however, will expose deeper layers of the deposit that likely contain initial abundances of fine-grained ash particles. The final stage of deposit erosion will expose a basal layer that has likely been enriched in ash particles through rainwater infiltration. These significant changes in tephra-fall deposit morphology and granulometry through time are poorly constrained.

Erosion of basaltic tephra-fall deposits through time can be constrained initially through examination of reasonably analogous deposits. Only trace amounts of tephra-fall deposit remain within 3 km of the roughly 80 ka Lathrop Wells volcano. Excluding deposits preserved in irregularities on associated lava flows, fall deposits have been completely eroded from other YMR volcanoes. In contrast, fall deposits are significantly intact 20 km from the vent at the 1065 A.D. Sunset Crater, Arizona (Amos, 1986), and the 2 ka Xitle volcano near Mexico City (Delgado-Granados, et al., 1998), both located in areas that receive 3–4 times YMR average rainfall. Although fall deposits are eroded within decades from areas with steep topographic gradients, deposits on relatively flat-lying areas are resistant to erosion (Segerstrom, 1960; Malin, et al., 1983; Inbar, et al., 1994). Based on comparison with these young analog deposits, staff conclude that tephra-fall deposits will likely be present up to 10,000 yr after deposition in the semiarid environment 20 km from the proposed repository site. In the TSPA-VA, DOE did not evaluate doses from contaminated tephra-fall deposits for times greater than 1 yr following the eruption (CRWMS M&O, 1998b). Current DOE models assume that volcanic tephra-fall deposits are <1 cm thick at the critical group location, and that soil removal rates for arid, nonvegetated agricultural areas are reasonable analogs for modeling tephra-deposit evolution (CRWMS M&O, 2000i). This approach does not appear realistic because it does not consider that basaltic fall deposits may be significantly >1 cm thick, will have grain-size characteristics different from common agricultural soils, and likely will be covered in significant amounts of vegetation. Deposit removal rates used in CRWMS M&O (2000i) may be inappropriately high and would significantly underestimate the longevity of tephra-fall deposits at the proposed critical group location.

The proposed critical group is located in an area that wind and water can deposit and erode HLW-contaminated tephra. Models that abstract tephra-deposit evolution through time must consider potentially significant contributions from remobilized tephra-fall deposits, in addition to removal of tephra by wind and water. For any future eruption through the proposed repository site, some amount of tephra will be deposited on slopes that are part of the Fortymile Wash drainage basin (Figure 34). By analogy with Parícutin volcano, Mexico, slopes with moderate-to-steep topographic gradients (i.e., 15–60 percent) will experience rapid removal of tephra-fall deposits through sheet, rill, and channel erosion (Segerstrom, 1950). For longer periods of time, lower-gradient topographic surfaces mantled by tephra-fall deposits also will be denuded through sheet and rill erosion. Sediment residence times in the confined channel of Fortymile Wash should be relatively short. Bed-load sediments will move down the main Fortymile Wash drainage during periods of high water flow. Just north of Highway 95, the main Fortymile Wash drainage morphology changes from a steep-sided channel to a broad, braided fan system. This location represents the point that significant long-term sediment deposition occurs within the Fortymile Wash drainage system. Sediment deposition and alluvial aggradation continues south into the Amargosa Desert and overlaps the general area proposed for the critical group location (Figure 34).

The potential significance of tephra remobilization by water to the critical group location is illustrated in Figure 34. A small-volume eruption (magma volume 1.3×10^6 m^3) was modeled using eruption conditions in Hill, et al. (1998) with the wind directed at the proposed critical group location. The eruption produced 2.9×10^6 m^3 of tephra, with about 1.4×10^6 m^3 (65 percent) of the tephra deposited on an erosional surface (Figure 34). Assuming the depositional area of the Fortymile Wash drainage basin is delineated by the mapped extent of surface drainages that extend south to the Amargosa River, remobilized sediment can be deposited over a 100 km^2 area. Assuming the remobilized tephra is deposited uniformly over this area, roughly 1 cm of remobilized tephra can be deposited through time at the proposed critical group location. This simplified analysis demonstrates that tephra remobilization by water could significantly affect the radionuclide concentration through time at the critical group location. Additional work is needed to evaluate basic assumptions regarding expected variations in tephra distribution due to the ambient wind field, characteristics and rates of tephra remobilization, and depositional patterns in the Fortymile Wash drainage system. This work must be accomplished before calculating the resulting impact on expected annual dose.

In addition to water, tephra can be remobilized by wind following the eruption. Near-surface winds are strongly affected by surface topography and, thus, have a significant north-south component (see Section 3.2.2.1.2). Tephra deposited in most directions from the critical group has the potential to be redistributed by wind to the critical group location. In addition, crops and fallow vegetation grown by the critical group likely will act as traps for wind-blown tephra (e.g., CRWMS M&O, 2000i). Relative to nearby, poorly vegetated areas, the critical group location likely has a higher potential for wind-blown sediment deposition through time. Models that account for tephra remobilization following the eruption also will need to account for remobilization effects by wind, in addition to water.

3.2.5.2 Summary

Calculations of expected annual dose will need to evaluate how the characteristics of basaltic tephra-fall deposits change through time in the arid surficial environment around Yucca Mountain. Inhalation of HLW-contaminated particles dominates current dose calculations. There are, however, only limited data on airborne particle concentrations above nonweathered basaltic tephra-fall deposits. These data indicate airborne particle concentrations may be significantly higher than reported for many poorly analogous deposits used in current performance calculations. In addition, there are no data on how these initial particle concentrations may change through time as the tephra-fall deposit is exposed to physical and chemical weathering processes. The proposed critical group also is located in an area that can receive a potentially significant influx of HLW-contaminated tephra through remobilization by wind and water, in addition to removal of tephra by these processes. Current models have not evaluated these potential remobilization effects through time. Calculations of annual risk to the proposed critical group will need to evaluate how the characteristics of tephra-fall deposits change through time, because most of the calculated risk is incurred through exposure to a tephra-fall deposit that endures for many years after the eruption event.

4.0 STATUS OF ISSUE RESOLUTION AT STAFF LEVEL

Consistent with NRC regulations on prelicensing consultations and a 1992 agreement with DOE, staff-level resolution can be achieved during prelicensing consultation. The purpose of issue resolution is to assure that sufficient information is available on an issue to enable the NRC to docket the license application. Resolution at the staff level does not preclude an issue being raised and considered during the licensing proceedings, nor does it prejudge what the staff evaluation of that issue will be after its licensing review. Issue resolution at the staff level during prelicensing is achieved when the staff has no further questions or comments at a point in time regarding how the DOE is addressing an issue. Pertinent additional information could raise new questions or comments regarding a previously resolved issue. Staff conclude the probability subissue is “closed-pending,” based on DOE agreements to evaluate risk using an annual probability of 10^{-7} for igneous events.

Many technical concerns in the consequences subissue have been addressed by the DOE since the TSPA-VA. The consequences subissue is nearing closed-pending resolution status. Significant amount of information are still needed from the DOE, however, to address staff concerns with models for magma-repository interactions and long-term surface remobilization effects. Although the consequences subissue remains “open,” staff are optimistic that information provided by the DOE in FY2001 will be adequate to move this subissue to a “closed-pending” resolution status. Details on subissue resolution are provided in the following sections. The basis for the status of subissue resolution focuses on data and models presented in this report and on agreements reached with the DOE during the August 2000 Technical Exchange on IA.

4.1 STATUS OF RESOLUTION OF THE PROBABILITY SUBISSUE

Prior to the August 2000 Technical Exchange with the DOE, staff had identified 12 specific technical concerns regarding the probability subissue. Details of these concerns are contained within Section 3 of this report, which follows U.S. Nuclear Regulatory Commission (1999). To address these concerns, the DOE agreed to resolve the probability subissue by providing in the Site Recommendation and License Application, in addition to DOE’s licensing case, the results of a single point sensitivity analysis for extrusive and intrusive igneous processes at an annual probability of 10^{-7}. By agreeing to provide these analyses, staff consider the probability subissue closed-pending, because the 10^{-7} analyses provide a reasonably conservative approach for evaluating risks from igneous activity.

Probability subissue resolution is not contingent on addressing the following staff technical concerns with existing DOE information, assuming that analyses using a 10^{-7} annual probability for igneous events are provided in relevant DOE documents. These staff concerns are provided, however, to give a summary of the range of technical concerns that the DOE will need to address in the absence of providing the 10^{-7} annual probability analyses.

1. **Recurrence Rates**. Discussions are insufficient in CRWMS M&O (2000b, j), because maps and other documentation do not indicate that all known or potential basaltic igneous features in YMR have been considered. DOE needs to demonstrate that igneous features are not present but undetected in the YMR. DOE has agreed to evaluate results of the new USGS Amargosa Aeromagnetic survey (Blakely,

et al., 2000) and present criteria used to determine if additional igneous events may be present but undetected in the YMR.

2. **Extent of Yucca Mountain Igneous System**. Although many different definitions are possible, DOE needs to provide a clear and consistent definition (e.g., CRWMS M&O, 2000j, figure 3-1, which lacks all buried igneous features). Definition of the YMR igneous system must be consistent with data used to define the igneous system in TSPA models and associated parameters.

3. **Stress/strain and Volcanism**: Discussions in CRWMS M&O (2000b, j) are insufficient to explain how recurrence rates may or may not change during the next 10,000 yr relative to long-term average recurrence rates. DOE needs to evaluate ongoing work by Wernicke, et al. (1998), Savage et al. (1998), and Dixon, et al. (2000).

4. **Miocene Basalt.** Ongoing work suggests Crater Flat Basin basalts since about 12 Ma may have a common petrogenesis, whereas 7–12 Ma YMR basalt petrogenesis may be strongly influenced by silicic caldera-forming processes. Miocene basalt in the Crater Flat basin, thus, provides relevant information for risk assessments not included in current DOE models.

5. **Literature Values**. Summarys in CRWMS M&O (2000b, j) are incomplete and do not adequately represent the preponderance of published literature since 1982 that indicates the annual probability of volcanic disruption of the proposed repository site ranges from about 10^{-8} to 10^{-6}. For example, Ho (1995) has an annual probability of volcanic disruption of 1×10^{-7} to 3×10^{-6}, and Ho (1992) ranges from 1×10^{-7} to 7×10^{-7}. The relationships between volcanic disruption probabilities in the literature and dike intersection probabilities in Geomatrix (1996) also should be clarified.

6. **Event Definitions**. Although generally clear in CRWMS M&O (2000b, j, k), Geomatrix (1996) used inconsistent definitions for extrusive and intrusive event types in calculating event probabilities. Different models or experts used different types of features to represent events in probability models. The resulting probabilities combine differing characteristics of extrusive and intrusive features. These differing models, however, are convolved into a singular probability distribution interpreted as solely representing intrusion intersection of the proposed repository site.

7. **Relationship of Source-Zones to Structure, Tectonic Models, and Geophysics**. Although some volcanic source-zones in Geomatrix (1996) and CRWMS M&O (2000b) are supported by tectonic models, many other zones and other tectonic models are not supported. Few tectonic models or data are cited in Geomatrix (1996) for zone definitions. Currently available geophysical data (gravity, aeromagnetic, and seismic) do not support zone definitions used in Geomatrix (1996) and CRWMS M&O (2000b).

8. **Effect of New Information on Elicitation**. Significant new geophysical (gravity, magnetic, and seismic) data, alternative tectonic models, and alternative

probability models have been developed since the 1995 PVHA elicitation. The extent of this information makes it likely that an expert's view of the igneous system now would be significantly different (cf. U.S. Nuclear Regulatory Commission, 1996). Parameter sensitivity studies (e.g., CRWMS M&O, 2000b) do not address this fundamental concern.

9. **Validation of Models**. Validity of PVHA source-zone modeling approach does not appear established by DOE. The DOE needs to demonstrate that its preferred approach can reasonably forecast the timing and location of future igneous events (cf. Condit and Connor, 1996).

10. **Elicitation Process**. Only a limited range of experts was selected by DOE for the PVHA using an internal nomination rather than a self-selection process. Potential biases or conflicts of interest to the experts are not documented. Modifications to initial elicitation reports also are not documented. These items do not follow the guidance in U.S. Nuclear Regulatory Commission (1996) for conduct of an expert elicitation, and, therefore, make it difficult to evaluate the conclusions of the PVHA elicitation (Geomatrix, 1996).

11. **Current Model Contradicts PVHA**. PVHA volcanic source-zones clearly were defined on timing and location of past volcanism within the source zone. A new "event center" (i.e., volcano) forms only in the source zone, with only a subsurface intrusion potentially extending out of the zone and intersecting the repository. The model in CRWMS M&O (2000b), however, has new volcanoes forming randomly along the dike, sometimes outside of the predefined volcanic source-zone. By PVHA definition, new volcanoes should occur only within the source-zone at recurrences defined by past patterns of activity within that zone. If volcanoes can form outside the source-zone as indicated in CRWMS M&O (2000b), the source-zones must be expanded to encompass the location of future volcanism. The frequency of dike intersections would then increase using the expanded zones, as shorter, more abundant dikes would intersect the proposed repository location.

12. **Use of PVHA Data**. Selective use of data from Geomatrix (1996) occurs in CRWMS M&O (2000b). For example, vent spacing (CRWMS M&O, 2000b, 6.5.2.2) only uses data from the 1 Ma Crater Flat and 0.3 Ma Sleeping Butte volcanoes but ignores relevant information from the 3.7 Ma Crater Flat, buried anomalies in Amargosa Desert, Paiute Ridge Intrusive Complex, and other features used to support igneous process models for the YMR. There also is an assumption that a relationship exists in Geomatrix (1996) between the number of events and the number of dikes. Geomatrix (1996) considered these as independent parameters.

4.2 STATUS OF RESOLUTION OF THE CONSEQUENCES SUBISSUE

Based on available information, staff conclude that basaltic volcanic eruptions characteristic of the YMR are capable of disrupting HLW canisters, entraining fragmented HLW, and dispersing this waste 20 km or greater downwind. There is considerable uncertainty in applying volcanological data and process models derived from undisturbed geologic settings to the engineered systems located in the disturbed geologic setting of the proposed repository site. Directed technical investigations still are needed to evaluate uncertainties associated with the entrainment and dispersal of HLW during volcanic eruptions, to determine granulometric characteristics of basaltic tephra-fall deposits through time, and to quantify interactions between basaltic magma, HLW, and waste canisters. Staff conclude, however, that conservative assumptions on available data provide a reasonable basis to conduct assessments of volcanic consequences on repository performance, with the understanding that these assessments may change substantially as new information becomes available. The NRC continues to review the data and assumptions inherent in its modeling to evaluate the degree of conservatism in the analysis and to reduce undue conservatism when warranted by the availability of data or models.

Significant changes were made to DOE igneous activity models following the TSPA-VA, as discussed in Section 3.2 of this report. These changes have addressed many of the staff's technical concerns with key modeling assumptions previously made by the DOE. Most importantly, the DOE currently assumes all HLW is available for entrainment and transport from waste packages that are intersected by an erupting volcanic conduit (CRWMS M&O, 2000d). In addition, there now is a significant reduction in HLW particle size during volcanic disruption, and all eruptions have violent strombolian dispersal characteristics (CRWMS M&O, 2000d). The consequence subissue remains open, however, due to significant uncertainties in how the DOE will evaluate magma-repository interaction processes (e.g., CRWMS M&O, 2000g) and long-term effects of remobilization (e.g., CRWMS M&O, 2000i). Discussions during the August 2000 technical exchange indicated the DOE will continue to evaluate these interaction processes, however, details on the modeling approaches will not be available until after submittal of the Total System Performance Assessment for Site Recommendation (TSPA-SR). In addition, wind and water may remobilize HLW away from or into the critical group location, which would affect long-term risk calculations significantly. Although the DOE will continue to evaluate remobilization processes, significant uncertainties will remain after submittal of the TSPA-SR. Staff are optimistic that the consequence subissue can be moved to a “closed-pending” status once DOE provides additional results of magma-repository interaction and remobilization studies.

Prior to the August 2000 technical exchange with the DOE, staff had identified 12 specific technical concerns regarding the consequences subissue. Details of these concerns are contained within section 3 of this report, which follows U.S. Nuclear Regulatory Commission (1999). The following agreements were reached with the DOE, which defined specific actions needed by the DOE to resolve these technical concerns.

1. **Magma/Repository Interactions**. If ascending magma intersects a drift, the extent and character of flow into the drift directly controls the number of waste packages disrupted. DOE will need to evaluate the extent and character of magma flow into a nonbackfilled repository, in addition to the backfilled drift scenario in CRWMS M&O (2000e). Models for conduit development (CRWMS M&O, 2000e, f), which controls the amount of HLW entrained in an eruption, will need to consider the effects of magma flow into drifts.

    - DOE will document the way in which the orientation of the repository drifts affects the number of waste packages incorporated into the volcanic conduit. Possible consequences of conduit elongation parallel to drifts will be documented in the TSPA-SR Revision 1, available to the NRC in June 2001.

- DOE will evaluate thermal and mechanical effects, as well as shock, in assessing the degree of waste package damage during magma flow into drifts. This evaluation will need to expand on current models to determine the extent and character of magma flow in drifts. DOE will document the results of these evaluations in interim change notices (ICN) to CRWMS M&O (2000b, d–f).

The process of magma-repository interactions is difficult to model numerically, and there are no known natural analogs. Staff continue to develop numerical and analog experimental models to evaluate results from DOE investigations, because uncertainties in this process affect risk calculations significantly. Due to the large uncertainties in current models, a limited amount of available data, and the relatively complex models still to be developed, staff view resolution of this concern as open. Staff will reevaluate the resolution of this concern following receipt of additional DOE reports in January 2001.

2. **Waste Package/Magma Interaction**. Current DOE models conclude that waste packages incorporated into an erupting volcanic conduit are wholly breached, and all HLW is available for entrainment (CRWMS M&O, 2000d, f). Staff agree with this conclusion (U.S. Nuclear Regulatory Commission, 1999). For intrusive events, DOE concludes up to three waste packages on either side of the intrusion will be wholly breached (CRWMS M&O, 2000f). Other waste packages further from the intrusion, however, have only limited end-cap damage. Staff are concerned that the primary damage zone will be more extensive than assumed by the DOE, and appropriate thermal and mechanical effects were not considered in defining the extent and character of the secondary damage zone.

- DOE will evaluate thermal and mechanical effects, as well as shock, in assessing the degree of waste package damage during magma flow into drifts. This evaluation will need to expand on current models to evaluate the extent and character of magma flow in drifts. DOE will document the results of these evaluations in interim change notices (ICN) to CRWMS M&O (2000b, d–f).

3. **Waste Form/Magma Interaction**. DOE concludes that thermal and mechanical effects during igneous events will reduce HLW particle sizes significantly (CRWMS M&O, 2000g). Minor differences between sizes in CRWMS M&O (2000g) and U.S. Nuclear Regulatory Commission (1999) need to be evaluated for significance. In addition, staff have ongoing concerns about how ash-particle density is modeled when HLW is incorporated into the ash.

- DOE will reexamine the ASHPLUME code to confirm that particle density is appropriately changed when waste particles are incorporated into the ash, and document the results in an ICN to CRWMS M&O (2000d). DOE also will conduct sensitivity studies on HLW particle sizes, which will be provided in Revision 1 to the TSPA-SR.

4. **Wind Characteristics**. Wind speeds need to be appropriate for modeling dispersal from 2 km to about 7 km high eruption columns characteristic of YMR volcanoes. Current DOE models use lower altitude wind speeds, which are relatively low velocity and, thus, have limited dispersal capability.

   - DOE will develop wind speed data appropriate for the height of the eruptive columns being modeled. These data will be documented in Revision 1 to the TSPA-SR.

   In addition to obtaining high-altitude wind speed data from the National Oceanographic and Atmospheric Administration (NOAA) for the Desert Rock airstrip, staff will evaluate the utility of a stratified wind-field in the ASHPLUME code. The stratified wind-field likely will provide a more realistic estimate of the dispersal capabilities for YMR basaltic volcanoes.

5. **Mass Loading Parameters**. Airborne particle concentrations above basaltic tephra-fall deposits need to be appropriate for the habits and lifestyles of the critical group. Few data are available for deposits reasonably analogous to basaltic tephra-fall deposits, and these data generally do not consider surface disturbing activity (CRWMS M&O, 2000l). In addition, most airborne particle concentrations do not consider 10–100 µm diameter particles, which are inhalable and have significant dose effects (see item number 7)

   - DOE will document the basis for airborne particle concentrations used in the TSPA-SR in Revision 1 to CRWMS M&O (2000l).

   Staff will complete technical investigations on grain-size characteristics of basaltic fall deposits and associated airborne particle concentrations. These data will be used to evaluate DOE models and data because relevant information is not available in the literature.

6. **Remobilization**. The HLW-contaminated tephra-fall deposit will be modified by wind and water for many years after the eruption. HLW can be transported away from and into the critical group location by wind and water following most future eruptions. These processes may result in a net increase in HLW through time, although this effect is poorly constrained. The long-term remobilization of HLW directly affects risk to the critical group, yet there are few data and models available to evaluate this process.

   - DOE will develop a linkage between the soil removal rate used in the TSPA-SR and the surface remobilization processes characteristics of the Yucca Mountain region, which includes additions and deletions to the system (CRWMS M&O, 2000i). DOE will document its approach to include uncertainty related to surface-redistribution processes in Revision 0 of the TSPA-SR. DOE will revisit the approach in Revision 1 of the TSPA-SR.

   Staff also will continue model development to evaluate remobilization processes, as remobilization is significant to risk calculations. In addition, several alternative modeling approaches are available, and uncertainties on data and models appear significant. Independent evaluation of remobilization, thus, is needed.

7. **Coarse Particle Inhalation**. DOE needs to use appropriate dose conversion factors (DCF) for 10–100 μm diameter particles, including direct absorption as well as ingestion effects.

    - DOE will provide additional justification on the reasonableness of the assumption that the inhalation of particles in the 10–100 μm range is treated as additional soil ingestion, or change the DCFs to reflect ICRP-30. The results will be documented in Revision 1 to CRWMS M&O (2000l).

8. **Self evacuation**. Previous DOE models proposed the critical group would self-evacuate in response to a basaltic eruption 20 km away. Staff concluded this was an unreasonable assumption, because historical basaltic cinder cone eruptions have not prompted self-evacuations 20 km away from the volcano. The current modeling approach (CRWMS M&O, 2000d, j) no longer assumes self-evacuation of the critical group, thus addressing staff's concern.

9. **Model for Airborne Transport**. The model used by the DOE to calculate the amount of HLW deposited at the critical group needs to be validated, that is, shown to reasonably calculate deposit characteristics of basaltic volcanic eruptions. DOE also needs to justify the model used for incorporation of HLW into the eruption, to ensure that HLW dispersal is calculated accurately.

    - DOE will document that the ASHPLUME model, as used in the DOE performance assessment, has been compared successfully with the 1995 Cerro Negro eruption (Hill, et al., 1998). This documentation will be provided in Revision 1 to the TSPA-SR. The model for HLW incorporation will be addressed in ICN to CRWMS M&O (2000d), as discussed in consequence item number 3.

    As shown in Hill, et al. (1998), staff conclude that the NRC ASHPLUME model reasonably calculates deposit thicknesses for basaltic volcanic eruptions. Staff will continue to evaluate HLW incorporation mechanisms and the effect of HLW entrainment on tephra dispersion to build confidence that the current approach is reasonable for evaluation of volcanic risks.

10. **Model for Groundwater Transport**. Although this concern largely is outside the scope of the IA KTI, preliminary results from DOE models suggest the effects of an igneous intrusion contribute to a larger component of risk in 10,000 yr than the effects of an extrusive volcanic eruption. In contrast, current NRC models indicate the risks from extrusive igneous events are substantially greater than those from intrusive igneous events. The basis for the different risk rankings is not clear, but will need to be evaluated.

- DOE will provide results in Revision 1 of the TSPA-SR that show the relative contributions of releases from primary and secondary waste-package damage zones resulting from intrusive igneous events.

11. **Integration of Results from All Pathways**. Staff was concerned with the method the DOE used to calculate the probability-weighted dose from igneous events. These concerns focused on the way that risk from prior year eruptions was combined with risk from an eruption in the year of interest. DOE clarified this methodology in the August 2000 technical exchange, indicating that all prior-year risk is being considered in risk calculations. This information was sufficient to resolve staff concerns.

12. **Volcano Type**. DOE needs to provide a technical basis for the volumes used to model future YMR volcanic eruptions (CRWMS M&O, 2000c, d). In addition, DOE needs to clarify that tephra volumes, and not entire eruption volumes, are used appropriately to model airborne dispersion.

    - DOE agreed and will document the basis for determining the range of tephra volumes that is likely from possible future volcanoes in the YMR in Revision 1 to the TSPA-SR, or demonstrate that TSPA-SR results are insensitive to uncertainties in the reasonably expected volumes of tephra in the YMR.

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Table 1. Example initial conditions and constants for eruption column model.

Initial Conditions	Example Value	Units	Explanation
n_o	0.01	dimensionless	mass fraction of gas at vent
L_o	10	m	vent radius
u_o	50	m s^{-1}	velocity at vent
θ_o	1100	K	temperature of column at vent
C_{po}	1617	J kg^{-1} K^{-1}	heat capacity of column at vent
R_{go}	462	J K kg^{-1}	molecular weight of gas in eruption column at vent × gas constant
Constants			
σ	1000	kg m^{-3}	density of solid pyroclasts
P	10000	Pa	air pressure
t	293	K	air temperature
C_a	998	J kg^{-1} K^{-1}	heat capacity of air

Table 2. Thermophysical properties used in heat transfer models, from Manteufel (1997) and McBirney (1984).

Physical Properties	Canister	Basaltic Magma	Wall Rock (Ignimbrite)
Thermal Conductivity (W/m °K)	50	1.25	2.1
Bulk Density (kg / m^3)	7800	3000	2200
Heat Capacity (J/kg °K)	450	1041	930
Thermal Diffusivity (m^2/s)	1.4×10^{-5}	4.4×10^{-7}	1.0×10^{-6}

Table 3. Volumes of historically active basaltic volcanoes used to estimate fall-deposit volumes for YMR Quaternary volcanoes.

Volcano	Age	Cone (km^3)	Lavas (km^3)	Falls (km^3)	Fall/cone	Fall/lava	Cone/lava
Tolbachik Cone 1	1975 A.D.	0.093	0.025	0.122	1.3	4.8	3.6
Tolbachik Cone 2	1975 A.D.	0.098	0.242	0.099	1.0	0.4	0.4
Sunset Crater	1200 A.D.	0.284	0.150	0.440	1.6	2.9	1.9
Parícutin	1943–1951 A.D.	0.069	0.700	0.410	5.9	0.6	0.1
Heimaey	1973 A.D.	0.015	0.180	0.012	0.8	0.1	0.1
Serra Gorda	<5 ka	0.030	0.015	0.042	1.4	2.8	2.0
Cerro Negro	1850–1995 A.D.	0.080	0.043	0.132	1.7	3.1	1.8
Lathrop Wells	0.13±0.01 Ma	0.024	0.038	0.048	2	n/a	0.6
Hidden Cone	0.38±0.02 Ma	0.019	0.009	0.038	2	n/a	2.0
Little Black Peak	0.31±0.02 Ma	0.006	0.007	0.012	2	n/a	0.9
SW Little Cone	0.90±0.02 Ma	0.002[a]	0.022	0.004	2	n/a	0.1
Red Cone	1.01±0.04 Ma	0.005[b]	0.089	0.005	1	n/a	0.1
Black Cone	0.94±0.03 Ma	0.011[b]	0.065	0.011	1	n/a	0.2

Note: (a) Cone volume corrected for 50% erosion; (b) cone volume corrected for 33% erosion.

Data sources: Tolbachik (Budnikov, et al., 1983); Sunset Crater (Amos, 1986); Parícutin (Segerstrom, 1950); Heimaey (Self, et al., 1974); Serra Gorda (Booth, et al., 1978); Cerro Negro (Hill, et al., 1998). YMR volcanoes from USGS 7.5' topographic map data.

Table 4. Summary of eruption parameters with calculated column heights and eruption powers for historically active basaltic volcanoes reasonably analogous to YMR volcanoes. *DRE* is dense rock equivalent (i.e., nonvesiculated). *Wilson* refers to the method of Wilson, et al. (1978), where magma density is 2600 kg m^{-3}, specific heat is 1100 J kg^{-1} $°K^{-1}$, a 1055 °K temperature change, and thermal efficiency of 0.7. *Walker* refers to the method of Walker, et al. (1984).

Volcano	Column height (km)	Eruption duration (s)	DRE volume (m^3)	Wilson, column height (km)	Wilson, power (W)	Walker, column height (km)
Heimaey 1973	2	2.2×10^6	5.2×10^6	2.2	4.9×10^9	2.1
Parícutin 1943	4–6	7.3×10^6	1.9×10^8	4.0	5.6×10^{10}	3.9
Tolbachik Cone 1 1975	6–10	1.2×10^6	6.0×10^7	4.7	1.0×10^{11}	4.5
Tolbachik Cone 2 1975	2–3	3.3×10^6	4.6×10^7	3.4	3.0×10^{10}	3.3
Cerro Negro 1947	4–6.5	6.6×10^4	1.1×10^7	6.3	3.5×10^{11}	6.2
Cerro Negro 1968	1–1.5	3.6×10^6	4.5×10^6	1.9	2.6×10^9	1.8
Cerro Negro 1971	6	6.0×10^5	1.4×10^7	3.9	4.9×10^{10}	3.8
Cerro Negro 1992	3–7	6.4×10^4	1.1×10^7	6.4	3.6×10^{11}	6.2
Cerro Negro 1995	2–2.5	3.5×10^5	1.3×10^6	2.4	7.9×10^9	2.4
Data derived from Heimaey (Self, et al., 1974); Parícutin (Segerstrom, 1950); Tolbachik (Budnikov, et al., 1983; Doubik and Hill, 1999); Cerro Negro (Hill, et al., 1998).						