Conceptual Model For Small-Volume Alkali Basalt Petrogenesis:
Implications For Volcanic Hazards At The Proposed Yucca Mountain
Nuclear Waste Repository

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10.1 Introduction

Today, 31 countries operate about 450 nuclear power reactors supplying electric power to nearly 1 billion people, about 15 percent of the world population. Nuclear reactors generate about 17 percent of global electric power needs and a number of industrialized countries depend on nuclear power for at least half of their electricity. In addition, approximately 30 nuclear power reactors are presently under construction worldwide [Macfarlane and Miller, 2007]. A comprehensive summary of the principles, practices, and prospects for nuclear energy may be found in Bodansky [1996]. Concerns regarding energy resource availability, climate change, air quality, and energy security imply a continuing demand for nuclear power in the world energy budget [Craig et al., 2001]. However, to date no country has solved the problem of long-term disposal or storage of nuclear waste. Without a long-term solution, the viability of nuclear energy as an increasingly significant contributor to power generation in the long-range future remains unclear. There is broad consensus that geologic disposal is the safest feasible long-term solution to high-level waste and spent fuel disposal. Although a number of countries have ongoing geologic repository research programs, there is presently no operational geologic repository for spent fuel or high-level waste on Earth. In the United States, where spent nuclear fuel and high-level waste amounts to nearly 50,000 metric tons, about 15 percent of the world total, implementation has proven to be challenging both technically and politically. Nuclear waste is currently stored on-site at existing nuclear power stations and at several temporary storage facilities. Permanent geologic disposal, like the siting of a nuclear power plant, requires careful site selection. For geologic disposal, lithologic formations that can isolate radioactive waste from the surrounding environment and biosphere at geologic timescales of order $10^4$–$10^6$ years are a minimum requirement [Macfarlane & Ewing, 2006]. Of particular importance in this regard are the nature, consequences, and probabilities of volcanic hazards that can potentially compromise public, environmental, and biospheric safety at long-term nuclear waste storage sites.
Yucca Mountain in Nevada, USA was identified in the early 1980’s as a potential geologic repository for nuclear waste. Yucca Mountain is made up of silicic volcanic tuff—rocks composed chiefly of pyroclastic flow and fall deposits. The proposed Yucca Mountain Repository (YMR) lies on the western boundary of the Nevada Test Site (NTS) within the Basin and Range geologic province [Zoback et al., 1981; Thompson & Burke, 1974]. This region is geologically active, with transtensional deformation manifested by faulting, related seismicity, high $^{3}$He/$^{4}$He anomalies indicative of mantle degassing [Kennedy and van Soest, 2007] and volcanic activity. Intense study of the seismicity, seismic hazards, geohydrology, petrophysics, structural, tectonic, and volcanic history of the Yucca Mountain region for over 30 years has provided the geologic foundation for the Total System Performance Assessment (TSPA) used by Yucca Mountain project geoscientists to make probabilistic forecasts of repository behavior. The TSPA considers all potential paths of radionuclides into the environment and defines the US Department of Energy’s (DOE) understanding of expected repository performance if built and operated according to present plans. The TSPA forms an integral part of the license application that the DOE will submit to the Nuclear Regulatory Commission (NRC) in mid 2008.

The consequences of magmatic disruption of the repository could be very significant in terms of the TSPA regarding radionuclide dispersion. An eruption beneath, into, or through the repository could lead to wide dispersal of radionuclides via atmospheric, surficial (i.e., particulate sedimentation), fluvial, and groundwater paths into the biosphere. Therefore it is important to evaluate the probability and consequences of potentially disruptive magmatic events. The consequences of disruption evidently depend on the characteristics of a magmatic event (e.g. eruptive style and volume, magma properties and dynamics, volatile content of magmas, etc). A disruptive igneous event within the footprint of the repository has been estimated to have a rather low occurrence probability of one event in 70 million years, with a 90 percent confidence interval ranging from 1 event in 20 million years to one event in 180 million years [Geomatrix Consultants, 1996]. Independent of the DOE, the NRC has estimated volcanically-induced
disruptive event probabilities in the range of 1 event in 10 million years to 1 event in 100 million years.

In this study we present a sketch of a conceptual model consistent with the eruption dynamics, petrology, major and trace element geochemistry of small-volume alkali basaltic volcanism in the Yucca Mountain region. In particular, the generation, modification upon ascent, and eruption of alkali basaltic magma to form small-volume volcanic constructs of lava and tephra is considered in some detail. This style of eruption, in our opinion, is the type most relevant to possible future eruptions at or near Yucca Mountain. Quantitative evaluation of volcanic hazards including eruption forecasting ultimately requires integration of comprehensive quantitative models of petrogenesis (*sensu lato*) into the regional geological framework. The aim of this study is to present a birds-eye-view of small-eruptive volume alkali basalt petrogenesis relevant to evaluation of magmatic hazards at the proposed permanent nuclear waste repository at Yucca Mountain, Nevada, USA.

### 10.2 Volcanic Hazard Evaluation

Evaluation of volcanic hazards at Yucca Mountain is aided by understanding the fundamental nature of regional volcanism. Results of previous geologic investigations of Yucca Mountain and environs clearly indicate a proclivity for small volume basalt eruptions, as opposed to, for example, large-volume, highly explosive silicic eruptions. However, evaluation of volcanic hazards at Yucca Mountain is complicated compared to traditional volcanic hazard studies that assess the possibility of future eruption at an existing volcano, because volcanism forecasting relevant to the Yucca Mountain Repository (YMR) involves an event at a specific location where volcanism has not occurred in the last 10 Ma. Volcanic hazard evaluation and prediction at Yucca Mountain therefore involves the coupled and difficult problems of predicting the timing, location, volume and eruptive style of possible eruptive events. Any single one of these issues can be complicated; together they represent a very challenging problem. In addition, high resolution
modeling of magmatic phenomena and eruption forecasting is itself complicated by the complex nature of dynamic processes in magmatic systems, which are non-linear at multiple spatial and temporal scales [Shaw, 1987], and the inherently stochastic distribution of heterogeneity at all scales in geologic media such as the upper crust beneath Yucca Mountain.

Information necessary for assessing the probability and consequences of a future disruptive magmatic event (volcanic eruption or intrusion by dike or sill at the repository depth) at Yucca Mountain includes the eruptive style(s) (e.g., pyroclastic flows or falls, lava flows, lahars, phreatomagmatic explosions, etc), spatial and temporal distribution, and number of previous events, as well as the distribution and geometric properties of the subsurface magma transport system (dikes, sills, hypabyssal intrusions, conduits). Assessing associated uncertainties is as important as defining mean values or average types of behavior. Constraints derive from, for example, careful study of analog volcanic provinces, including older, exhumed terrains, in the context of the regional geologic and geophysical setting, as well as application of dynamical, phase equilibria, and trace element models relevant to magma generation, transport, reaction, and eruption based on thermodynamics, fracture mechanics, fluid dynamics and geochemistry.

This paper is organized as follows. In the next section, a brief summary of Yucca Mountain region volcanism is given. This provides a basis for inferring the most likely composition, volume, volatile content, and eruptive style of volcanic events in the next ten thousand to one million years. Based on this determination, we consider stages in magma transport from source to surface. These stages include the phase equilibria and thermodynamics of partial melting, segregation, mobilization and ascent of magma through the lithosphere, and finally, near-surface flow of magma, driven mainly by volatile exsolution.

10.2.1 Volcanological And Tectonic Background

Over the past 30 years, many studies have addressed the age, geochemistry, petrology, volcanology and magma properties relevant to magmatic activity in the greater Yucca Mountain
region. It is beyond the scope of this work to review this literature. A few recent studies that serve as links to earlier work include Smith et al. [1990], Fleck et al. [1996], Perry et al. [1998], Valentine et al. [1998], Perry and Youngs [2000], and Smith et al. [2002]. The report of Detournay et al. [2003] summarizes the history of volcanism, describes magma thermodynamic and transport properties, and addresses the likely characteristics of a future volcanic event at Yucca Mountain based on past volcanic activity, with a focus upon the past 5 million years. The consequences of repository disruption by an igneous event are discussed in Detournay et al. [2003] and Woods et al. [2002]. Crowe et al. [2006] present an overview of the volcanism problem, how it has been studied historically in the context of the Yucca Mountain project, and its potential impact on an underground repository. The study of Fridrich et al. [1999] provides a thorough account of the tectonics, especially the evidence and timing (mainly Miocene) of extension in the Yucca Mountain region. Regional volcanism is summarized in Perry et al. [1998]. Lathrop Wells, the youngest volcano in the region is described in detail by Perry and Straub [1996], Valentine and Perry [2007] and Valentine et al. [2006]. We draw upon these studies in the summary below.

A regional map that highlights recent volcanism around Yucca Mountain is given in Figure 10.1. Yucca Mountain (sensu stricto) is composed of Miocene silicic volcanic rocks representing deposits from a series of large-volume eruptions associated with several large calderas north of the mountain. The most proximate one is the Timber Mountain Caldera north of Yucca Mountain. Silicic pyroclastic eruptive activity began about fifteen million years ago and ceased about eight million years ago. Patterns of silicic ignimbrite-forming events in the Great Basin over the past thirty million years indicate that further large-volume silicic pyroclastic flow and fall eruptions are not likely to recur in the Yucca Mountain region within the next few million years. Late Miocene-Quaternary basaltic volcanic activity succeeded Miocene silicic volcanism in the Yucca Mountain region. The basalts can be divided into two major episodes. Basalt of the silicic episode erupted during the waning stage of silicic volcanism (>8 Ma), whereas eruptions of
post-caldera basalt began more recently and continue into the Quaternary. The post-caldera episode can be further subdivided into older post-caldera basalts outcropping north, northeast, and southeast of Yucca Mountain (Figure 10.1), and younger post-caldera basalts, which crop out west, southwest, and south of Yucca Mountain. The older post-caldera basalts range in age from about nine million years before present to about six million years before present. An apparent volcanic hiatus of about two and a half million years (roughly from 7.2 to 4.7 Ma) separates the older and younger post-caldera basalts.

In order of decreasing age, the younger (Pliocene–Quaternary) basaltic volcanics (lava flows and tephra deposits) include the Pliocene basalts of Thirsty Mesa (4.7 Ma; volume, V~2.6 km$^3$) and southeast Crater Flat (3.7 Ma; V~0.6 km$^3$), the 3.1 Ma basaltic trachyandesites of Buckboard Mesa (V ~0.8 km$^3$), the five Quaternary (~1 Ma) alkali basalt cones of Crater Flat of total volume ~0.15 km$^3$, the 0.35 Ma basalt cones of Hidden Cone and Little Black Peak at Sleeping Butte (V~0.05 km$^3$), and the most recent eruption in the area, the 78 ka, ~0.1 km$^3$ Lathrop Wells alkali basalt cone and lava field located about 20 km south of the YMR footprint. Within the last million years, eruptions near Yucca Mountain have all been small volume; deposits include both tephra fallout and lava flows of alkali basalt (nepheline normative) composition.

In addition to exposed volcanic deposits, a number of buried basaltic lava flows or small intrusive bodies inferred from geomagnetic surveys lie beneath alluvial fan deposits in the shallow subsurface southwest (Crater Flat) and southeast (Jackass Flats) of Yucca Mountain (Figure 10.1). Drilling and geochronological, geochemical and petrographic examination of these bodies has been conducted at (1) the 11.3 Ma basalt of magnetic anomalies Q (R and 4, probably related to Q) and T located northwest of the ~1.1 Ma Red and Black cones in Crater Flat, east of the Bear Mountain Fault at the western edge of the Amargosa trough, (2) anomaly A, a 10.1 Ma basanite located due south of anomaly T and the Quaternary Little Cones, (3) a large-area magnetic anomaly in Jackass Flats of age ~9.5 Ma mainly south and east of Yucca Mountain forming the eastern boundary of the Amargosa Trough, and (4) the buried basalts of anomalies V
(9.5 Ma), B (3.8 Ma), and G (F and H probably related to G) dated at 3.8 Ma. Several additional magnetic anomalies inferred to be buried volcanics based on their locations and magnetic signatures have not been sampled (e.g., anomalies C and D). Although anomalies C and D have not been dated radiometrically, estimates based on depth of burial and alluvial fan sedimentation rates suggest an age between 4.2 and 5.8 Ma.

Except for the basaltic andesite of Buckboard Mesa and the basalt of anomaly B, all younger post-caldera exposed basalts lie within a 10-km wide northwest-trending zone west of Yucca Mountain in the Crater Flat–Amargosa trough corridor. This zone extends from the small-volume basalts of Sleeping Butte south through the Quaternary and Pliocene basalts of Crater Flat and continues south to Lathrop Wells and buried Pliocene basaltic deposits in the Amargosa Valley.

10.2.2 ‘Expected’ Event: Composition, Volume And Style

Discrete events in the region in the last few million years are characterized by small-volume (0.01 km$^3$ to ≤1 km$^3$) alkali basaltic lava and tephra eruptions. Based on modern analogs, typical eruption durations are measured in terms of days, weeks, or months and depend primarily on the total volume of magma erupted. Eruptive fluxes in the range $10^3$-$10^4$ kg/m$^2$ s lead to mass flows of order $10^5$-$10^7$ kg/s based on typical conduit cross-sectional areas observed in ancient analog systems [e.g., Detournay et al, 2004]. Eruptions at the high end of these ranges could produce Strombolian eruption plumes up to ten kilometers high; more typical eruptions would produce plume heights of order several kilometers. The volatile content, inferred from phase equilibria, the study of glass inclusions in phenocrysts, and the order of low-pressure, near-surface crystallization of microphenocrysts, lies in the range 2.5–4 wt. %, with H$_2$O the dominant fluid species [Detournay et al, 2004]. The dissolved volatile content of rising magma plays a critical role in determining eruptive style. A dynamical transition occurs when the volume fraction of the fluid phase in magma exceeds the critical volume fraction ($\theta_{cr}$) for magma fragmentation of circa 0.5–0.7. There is a rapid decrease in magma density and increase in
magma (mixture) compressibility around this rheological transition. These magma property variations lead to rapid increases in magma eruptive velocity near the fragmentation depth. The depth at which magma fragmentation occurs depends upon the dissolved volatile content of magma and the dependence of volatile solubility on temperature, pressure and melt composition. It also depends on whether or not volatiles can leak from the magma into surrounding crustal rocks. These issues are quantified in section 10.6.3. Although not discussed here, phreatomagmatic eruptions, in which magma encounters low-temperature, water-saturated permeable crust to generate stream-rich violent eruptions, is a possible eruption style in wetter climates than now observed at Yucca Mountain. Musgrove and Schrag [2006] have analyzed possible future climates, specifically wetter conditions associated with a higher water table, in the southern Great Basin. They noted that the most recent time in earth history when CO$_2$ levels approached those anticipated in the next few hundred years was in the Eocene. Consideration of Eocene-like climate scenarios may therefore provide some lessons about possible climate changes due to increased (anthropogenic) CO$_2$. The design of the repository should account for the potential risk of significantly wetter conditions at Yucca Mountain that would enhance the chances for phreatomagmatic style eruptions. In the remainder of this paper, phreatomagmatic eruptions are not explicitly considered.

10.3 Magma Generation And Transport: A Source To Surface Overview

We consider several topics related to the generation, segregation, ascent, and eruption of basaltic magma and consider the results in light of volcanic hazards at Yucca Mountain. We adopt as typical of future eruptive activity in the Yucca Mountain region the 78 ka, ~0.1 km$^3$ alkali basaltic lava and tephra eruption at Lathrop Wells, about 20 km south of the proposed YMR. Although the generation and transport of magma is a continuum process, it is convenient to analyze successive stages from melt generation to eruption or shallow level intrusion. We use the self-consistent, thermodynamically based pMELTS and MELTS phase equilibria models of Ghiorso et al. [2002] to perform calculations that account for important sources of variability in
liquid compositions and their physical properties. Extensive documentation of the phase equilibria algorithms is presented elsewhere [e.g., Hirschmann et al., 1998; Hirschmann et al., 1999a, 1999b: Ghiorso et al., 2002; Asimow and Longhi, 2004]. Although no phase equilibria model is perfect due to the thermodynamic complexities of multicomponent-multiphase silicate systems, the MELTS algorithm has been repeatedly shown to faithfully capture multicomponent phase relations in mafic-ultramafic systems. Petrogenesis from phase equilibria includes forward modeling of primary melt generation via partial melting (PM) of a peridotitic source and primary melt modification by fractional crystallization (FC) during upward ascent of primary melt. In particular, we compare melt compositions resulting from forward modeling of PM and FC to the available volcanologic, geochemical and petrologic database for Lathrop Wells [Perry and Straub, 1996; Valentine et al., 2006; Valentine and Perry, 2007]. The phase equilibria results link heat transfer between magma and surrounding lithosphere to the fracture mechanisms that drive magma ascent beneath Yucca Mountain, where the large-scale extensional stress environment of the Great Basin aids buoyancy-driven upward propagation of magma-filled crack networks. As magma ascends to the near surface, new dynamic processes become important. Once a magmatic mixture develops an appreciable volume fraction of fluid (bubbles), the behavior of the compressible dynamics of magma (melt plus fluid) becomes important. The dispersal of ash and lava from a small-volume alkali basalt volcano is very relevant to magmatic hazard analysis at Yucca Mountain.

The initial step leading to small-volume alkali basaltic eruption is partial melting of peridotite in the upper mantle. Factors that govern the composition of primary melt include the extent to which partial melt stays in chemical potential equilibrium with the crystalline residue, the composition of source peridotite (i.e., mineral compositions and abundances, water content, and redox state), and the mean pressure of PM. We tested the effects of each of these parameters by performing approximately eighty partial melting simulations systematically varying the governing parameters. Below, results for representative cases to illustrate parameter sensitivity
are presented. We then analyze in detail the melting scenario that produces primary partial melt, which upon further evolution by fractional crystallization, exhibits a composition corresponding most closely to Lathrop Wells basalt. Liquid ferric iron to ferrous iron ratios are based on the oxygen fugacity constraint along the QFM-1 buffer (i.e., one log base ten unit below the Quartz–Fayalite–Magnetite buffer). We performed a number of sensitivity tests to explore the effects of varying oxygen fugacity and found the effects to be minor compared to variations of other intensive parameters such as pressure or source fertility.

10.3.1 Magma Generation: Source Bulk Compositions Studied

Table 10.1 gives the anhydrous major element compositions of three end-member ultramafic compositions used in the phase equilibria simulations. The compositions are peridotitic, but differ in terms of fertility. Fertility refers to the potential of a peridotite to generate basaltic liquid (melt) by partial melting. There is no strict definition of fertility; several measures of fertility are given in Table 10.2. In particular, the mass ratio of the phases amphibole + phlogopite + clinopyroxene + garnet relative to olivine + orthopyroxene at the solidus pressure, modal clinopyroxene, the mass ratio of FeO/MgO in the peridotite source, and relatively high values for Al$_2$O$_3$, TiO$_2$ and CaO all positively correlate with peridotite fertility. Composition 1B-33 [Bergmann, 1982] corresponds to a depleted harzburgite xenolith from a 38±10 ka lava flow within the Lunar Crater Volcanic Field [Yogadinski, 1996], located several hundred kilometers northeast of Yucca Mountain. Composition 1B-33 represents the most depleted of the compositions used in this study. PA-12 [Frey & Prinz, 1978] is a moderately fertile peridotite xenolith from the Pleistocene basaltic vent at Peridot Mesa in the San Carlos Volcanic Field, Arizona, U.S.A. The most fertile peridotite composition used in this study, KLB-1 [Hirose & Kushiro, 1993], is a garnet peridotite xenolith from Kilbourne Hole, New Mexico, U.S.A. [Padovani & Reid, 1989]. The Kilbourne Hole maar, a phreatomagmatic explosion crater, formed when the Afton basalt intermingled with wet rift-fill sediments in the Camp Rice Formation of
the Santa Fe Group. The age of the phreatomagmatic explosion and basalt eruption responsible for creating the Kilbourne Hole maar is 77 ka [Anthony and Poths 1992].

10.3.2 Magma Generation: Fractional or Batch melting?

There are two end-member models relevant to PM. In batch partial melting, melt remains in chemical potential equilibrium with crystalline residue during progressive fusion until melt is segregated from source crystals at some fixed value of the extent of melting, either 5, 10 or 15 % melting (by mass) in this study. In fractional partial melting, melt is isolated from crystalline residue immediately and completely upon generation; no further reaction occurs between melt and residual crystals. In reality, the mode of PM in the Earth lies between these limits. We compared results of fractional and batch PM for all three starting compositions. Figure 10.2a-e shows MgO variation diagrams for the compositional sequence of partial melts generated by fractional and batch PM for starting composition KLB-1 (fertile peridotite; Table 10.1) plus 0.2 wt. % H₂O. Oxygen fugacity is constrained at QFM-1 and the pressure of isobaric melting is 1.5 GPa, corresponding to a depth of about 50 km beneath Yucca Mountain. The solidus is marked Tₛ and points are labeled on the liquid curves representing 5, 10 and 15 % partial melting by mass. The melt composition at the solidus is identical for batch and fractional PM. The variation in partial melt composition with extent of melting is largest at low degrees of melting in both cases, although the spectrum of melt compositions is greater in fractional PM than in batch PM, as expected. At larger degrees of melting, predicted liquid major element compositions differ modestly for silica, FeO, Al₂O₃ and CaO, and are most marked for the total alkalies (Na₂O + K₂O). In section 10.4 we present FC calculation results indicating that batch PM better represents mantle PM than fractional PM. In short, FC models based on melts derived from fractional PM produce liquids that deviate widely from observed compositions at Lathrop Wells, whereas batch PM generates liquids that, on undergoing FC, coincide quite well with observed Lathrop Wells major and trace element data. Based on trace element arguments, it is generally argued that the degree of melting (by mass) of a peridotite source to produce basalt is in the range 5–10 %. All
results reported on hereafter refer to batch PM with extent of melting between zero and 15 percent.

### 10.3.3 Magma Generation: Role Of Peridotite Fertility

We investigate the importance of peridotite fertility via PM computations based on the three peridotites of Table 10.1. 0.2 wt. % H$_2$O has been added to each anhydrous bulk composition. Table 10.2 gives the solidus phase assemblage for each bulk composition, calculated by Gibbs energy minimization at 1.5 and 3.5 GPa. Small amounts of modal amphibole and phlogopite (hydrous phases) are present at the 1.5 GPa solidus for the more fertile compositions PA-12 and KLB-1. Figure 10.3a-e depicts the composition of melts generated by batch PM at 1.5 and 3.5 GPa on MgO variation diagrams. The symbols along each curve indicate 5, 10, and 15 % melting by mass. The composition of melt at the solidus is marked T$_s$. The Lathrop Wells basalt compositional field, shown as a shaded field, is for reference only. Since Lathrop Wells basalt is *not* primary melt (its Mg# is far too low), calculated partial melt compositions are *not* expected to cross the Lathrop Wells field. However, because we expect that primary melt evolves to Lathrop Wells melt by fractional crystallization, the starting composition along the batch PM curve must lie at a location in composition space such that FC of the primary melt drives the liquid composition into the Lathrop Wells field. We demonstrate in section 10.4.1 that a fertile (KLB-1) or moderately fertile (PA-12) source undergoing PM at relatively low pressure (1.5–2 GPa) does indeed provide a composition, which upon subsequent FC during ascent, evolves into the Lathrop Wells compositional field. This is especially clear on the CaO and total alkalies MgO variation diagrams. The batch PM generated by 5, 10 or even 15 % PM of a depleted source cannot evolve by FC into the Lathrop Wells field. In addition, batch PM melting at 3.5 GPa generates liquids that do not, upon subsequent crystal fractionation, evolve into the Lathrop Wells compositional field. We conclude that the mantle source for Lathrop Wells basalt was fertile or moderately fertile and that the mean pressure of PM was significantly closer
to 1.5 GPa (~50 km) than 3.5 GPa (110 km). The effects of melting pressure are examined in more detail below.

10.3.4 Magma Generation: Role Of Pressure On Composition Of Partial Melting

The role of pressure in modifying the composition of peridotite batch partial melts has been studied in detail. The starting composition is fertile KLB-1 plus 0.2 wt. % H₂O. The redox condition is fixed along the QFM-1 buffer. MgO variation diagrams showing the compositional sequence of batch partial melts starting at the solidus are shown in Figure 10.4a-e for pressures of 1.5, 2.0, 2.5, 3.0 and 3.5 GPa. These pressures span a depth range from ~50 to 110 km beneath Yucca Mountain. The effect of increasing pressure at fixed melt fraction (e.g., 5 wt. %) is to increase MgO and FeO and reduce the Al₂O₃ content of partial melt. An important result of these calculations is that high-pressure PM can be ruled out. For example, the FeO vs MgO plot (Figure 10.4b) shows that partial melts generated at pressures less than or equal to 2 GPa are less FeO rich than high-pressure melts at the same extent of melting. Because FC of basaltic melt drives derivative liquids to higher FeO, it is clear that FC of a high-pressure partial melt cannot drive a primary melt into the Lathrop Wells field. The conclusion is that the pressure of PM for generation of the primary Lathrop Wells basalt liquid is in the range 1.5–2 GPa (~52–67 km beneath Yucca Mountain).

10.3.5 Magma Generation: Role Of H₂O Content In Source Peridotite

We performed a series of hydrous melting calculations based on a variety of source H₂O contents to explore the effect of H₂O on melting of fertile peridotite KLB-1. Figure 10.5a-e shows the calculated compositional paths for batch melting at 1.5 GPa, with oxygen fugacity at QFM-1. The plotted results correspond to four distinct source water contents: dry, 0.1 wt. % H₂O, 0.2 wt. % H₂O and 0.6 wt. % H₂O. The symbol boxes along the melt lines refer to 5, 10 and 15 % melting. The solidus is marked by Tₛ. An initial H₂O concentration of 0.6 wt. % exceeds the
storage capacity of melt and hydrous minerals (amphibole and phlogopite) and leads to a hydrous fluid phase at the solidus, a somewhat unlikely condition in general given inferences of the average water content of typical mantle peridotites [Carlson and Miller, 2003; Asimow and Langmuir, 2003]. For a fixed fraction of partial melt (e.g., 5 %) the alkali–silica diagram (Fig 10.5a) shows that partial melts become increasingly silica-rich (from ~46 wt. % to ~53 wt. % SiO₂) with increasing initial H₂O content, although the total alkali content of the melt remains constant around 4 wt. %. The criterion that pMELTS liquid H₂O concentration predictions at the end of fractional crystallization must coincide with inferred pre-eruptive Lathrop Wells H₂O concentrations of 2.5–4 wt. % (Detournay et al., 2004) leads to the conclusion that source water contents lie in the range 0.1–0.3 wt. %. We have chosen to present pMELTS results based on a representative starting H₂O concentration of 0.2 wt. %. Dry partial melts are too alumina and silica poor and too FeO rich to evolve by FC into the Lathrop Wells field.

**10.3.6 Magma Generation: Summary Of Partial Melting Conditions**

We have conducted over 100 phase equilibria simulations using the pMELTS algorithm of (Ghiorso et al., 2002) in which systematic variation of several critical factors (melting process, source fertility, pressure of melting, and source H₂O content) have been studied. Evaluation of the quality of results is based on the criterion that fractional crystallization of primary partial melts of a fertile peridotitic source should give rise to a predicted melt composition similar to that observed for Lathrop Wells basalt. We have identified the most plausible melting scenario as 5 to 10 % batch melting of a fertile peridotite (olivine, orthopyroxene, clinopyroxene, spinel, amphibole, and phlogopite) with an initial water content of ~0.2 wt. % H₂O along the QFM-1 buffer curve at 1.5–2 GPa (~50–70 km depth). At either 5 or 10 % melting, the residual source assemblage is olivine, orthopyroxene, clinopyroxene and spinel. The temperatures and dissolved H₂O contents of primary melt generated at 5 % and 10 % melting are 1280 °C and 1350 °C, and 3.3 wt. % H₂O and 1.74 wt. % H₂O, respectively at 1.5 GPa. The Mg# (≡ atomic Mg/(Mg+Fe²⁺))
of the 5 and 10 % partial melts are 76 and 77, respectively. The mineral proportion diagram for
batch PM of fertile peridotite KLB-1 is shown in Figure 10.6a. The solidus at 1.5 GPa is 1120 °C;
10 % batch melting is reached at ~1300 °C. The residual phase assemblage is olivine,
orthopyroxene, spinel and clinopyroxene. All modal amphibole and phlogopite are consumed
during PM within ~30 °C of the solidus.

10.4 Fractional Crystallization During Ascent

The Mg# of Lathrop Wells basalt is 54±2 [Perry & Straub, 1996]. Given that primary
liquid generated by PM of peridotitic sources has Mg# ~75–78, we conclude that Lathrop Wells
basalt is not primary melt. Assuming that Lathrop Wells basaltic melt originated by peridotite PM
followed by fractional crystallization during ascent, the mean pressure of FC can be estimated
based on phase equilibria constraints. Our approach involves using the primary melt products of
peridotite batch PM as input compositions for calculating FC compositional paths using the
MELTS and pMELTS algorithms [Ghiorso and Sack, 1995; Ghiorso et al., 2002]. The FC
calculations are ended when the Mg# of the computed melt is 54±2, the range observed for
Lathrop Wells basalt. We performed FC calculations in which the source fertility, mean pressure
(depth) of fractional crystallization, and fraction of primary partial melt (and therefore, primary
melt composition) vary systematically. We performed numerous FC calculations based on
primary melt fractions of 5, 10, and 15 %, but include here only FC results based on primary melt
fractions of 10 %, as these coincide most closely with observed Lathrop Wells major element
data. In each FC calculation, the initial water content corresponds to that of the primary melt
product of batch PM. The quality of the results is judged by (1) comparison of calculated melt
composition oxides SiO₂, Al₂O₃, FeO, CaO, K₂O and Na₂O with Lathrop Wells basalt at a
computed melt Mg# of 54 and (2) comparison of the computed H₂O content of derivative melt
(Mg# = 54) with the inferred water content of Lathrop Wells eruptive basalts of 2.5–4 wt. % H₂O
[Detournay et al., 2004]. In a later section, we use independent methods based on Lathrop Wells trace element data to test the validity of the phase equilibria constraints. In particular, we show that a mean pressure of fractional crystallization ~1 GPa (depth ~36 km) best fits observed trace element abundances. This depth lies in the shallow upper mantle beneath Yucca Mountain.

### 10.4.1 Fractional Crystallization: Mean Pressure and Implied Phase Relations

MgO variation diagrams are shown in Figure 10.7a-e for isobaric FC of primary melts formed by 10 % PM of fertile peridotite KLB-1 (Table 10.1) at 3.5 GPa and 1.5 GPa. We performed a sequence of FC calculations at successively higher pressure with maximum pressure of 3.5 GPa. Predicted major element concentrations do not compare well with those measured on Lathrop Wells basalts [Perry & Straub, 1996; Valentine and Perry, 2007] for FC at pressures above about 2 GPa. Here, we show isobaric FC paths at both 1 GPa and 0.5 GPa. The major feature is that FC of primary partial melts formed at high-pressure generates evolved melts that are too rich in FeO, CaO, and the alkalies, and too depleted in silica and alumina to match Lathrop Wells basalt. In contrast, partial melts generated at low pressure (1.5 GPa) evolve into the Lathrop Wells field during FC at pressures in the range 0.5-1.0 GPa. Liquids resulting from FC at 1 GPa of primary melt generated at 1.5 GPa better coincide with Lathrop Wells compositions than do those resulting from lower pressure (0.5 GPa) FC of the same primary melt.

The fractionating phases at 1 GPa are olivine, clinopyroxene and spinel. At 0.5 GPa, the fractionating phases are olivine, plagioclase, and trace clinopyroxene. The contrasting roles of plagioclase (low-pressure) versus clinopyroxene (high-pressure) fractionation account for the CaO–MgO variation diagram differences in Figure 10.7. In each case, the initial dissolved water content of the primary melt is 1.75 wt. % H\textsubscript{2}O at the start of FC. The dissolved water content of liquid at the conclusion of FC (melt Mg# ~52–54) is 3.2 wt. % H\textsubscript{2}O at 1.0 GPa and 2.6 wt. % H\textsubscript{2}O at 0.5 GPa, within the 2.5–4.0 wt. % H\textsubscript{2}O range of Lathrop Wells basalt pre-eruptive H\textsubscript{2}O contents estimated by Detournay et al. [2004]. If this exercise is repeated using 5 % partial melt
as the starting liquid for otherwise identical conditions, the liquid line of descent departs significantly from the Lathrop Wells field at both 1 GPa and 0.5 GPa. This is because the crystallizing phases of the 5 % partial melt are dominated by olivine and amphibole. Amphibole is not found as a phenocryst in Lathrop Wells basalts.

10.4.2 Fractional Crystallization: Summary Of Solidification Conditions

The composition of the Lathrop Wells basalts has been used to define an approximate set of compositional and geophysical parameters relevant to PM and FC based on phase equilibria modeling. The representative model suggests batch partial melting at ~50–60 km of fertile peridotite containing ~0.2 wt. % of H₂O, with redox conditions near the QFM-1 buffer. The extent of melting is in the range 7–10 %, perhaps closer to 10 % by mass. Magma ascends and undergoes fractional crystallization at a mean pressure of ~1 GPa (depth equivalent to 35–40 km) in the upper mantle. Because the extent of PM is about 10 % and about half the primary melt freezes at depth, the volume of contributing source zone mantle can be estimated. The eruptive volume (dense rock equivalent of lava plus tephra) of Lathrop Wells is approximately 0.12 km³. Allowing for the effects of distal ash dispersal and the effects of solidification during ascent (roughly 50% by mass), the total volume of melt generated by partial melting in the source is liberally estimated to be ~ 0.5 km³. If the extent of partial melting was about 10 wt. %, then the mantle melting volume was about 5 km³. This suggests a small ‘melting’ footprint compared to, for instance, the map distance between the southernmost and northern most basaltic cones of the Quaternary in Crater Flats (Figure 10.1) of about 15 km. For example, if the melting region is assumed spherical, then the radius of the melting region is ~ 1 km.

10.5 Trace Element Geochemistry: A Test Of The Phase Equilibria Model

It is important to independently test the conclusions of phase equilibria modeling. One test is provided by comparison of trace element concentrations in Lathrop Wells basalts with
predictions based on trace element modeling using the phase proportions and compositions taken from the phase equilibria model. Once the trace element abundances in the bulk source and mineral–melt partition coefficients are fixed, then the consequences of PM and FC can be forward modeled by numerical solution to the equations for batch PM and subsequent FC. The trace element models are based upon the phase assemblages computed from MELTS self-consistently [Spera et al., 2007]. Source trace element abundances used in the calculations are collected in Table 10.3. The values of the mineral–melt trace element partition coefficients are given in Table 10.4. These data together with phase assemblages from the phase equilibria modeling were used to compute the trace element contents of partial melts by batch melting followed by FC. The absolute trace element abundance pattern for Lathrop Wells basalts is portrayed as a band in Figure 10.8. The starting melt composition is the point marked T1. The cross marks the spot where the melt has a Mg# of 52. Recall that Lathrop Wells basalts have Mg# in the range 52–54. Twenty-three trace element absolute abundances have been calculated including representatives from the transition metals (e.g., Ni, Co, Cr, V), rare earth elements (La through Lu), large-ion lithophile elements (e.g., Rb, Ba, Sr, Pb) and the high-field-strength elements (e.g., Nb, Ta, Zr, Hf) among others. Overall the computed results agree well with abundances observed for Lathrop Wells basalts [Perry & Straub, 1996; Valentine & Perry, 2007]. We conclude that trace element modeling based upon the major element phase equilibria results are mutually consistent.

10.6 Ascent of Magma: Overview

One can distinguish several dynamical regimes relevant to the segregation and upward transport of magma associated with small-volume alkali basalt lava/tephra eruptions of the Lathrop Wells type. In the source region at depths 50-60 km, percolative flow (i.e., Darcy flow) of melt in response to small pressure gradients represents the first stage in the segregation of melt from residual crystals into networks of contiguous volumetric domains. If the melt generation rate
exceeds the rate at which ductile flow can accommodate the increased volume associated with melting, magma pressure will rise above the prevailing mean normal stress, $\sigma_n = (\sigma_1 + \sigma_2 + \sigma_3)/3$ and crack propagation will be favored. $\sigma_3$, the minimum principal stress is horizontal to sub-horizontal for regional transtensional stress states and therefore vertical to sub-vertical magma-pressured cracks could open. Melt slightly above the mean normal pressure can then segregate from crystalline residue and begin ascent under the combined forces of buoyancy and small (~few MPa) principal stress differences. Swarms of melt-filled cracks propagating upwards transport melt at rates depending on factors such as the fracture width, fracture resistance, the driving pressure gradient and melt viscosity. It is during this stage that fractional crystallization, probably along crack margins occurs as melt fractures rise through cooler lithospheric mantle. Rates of ascent in this stage for alkali basalts are in the range cm/s to m/s [Spera, 1986]. Alkali basalts moving at the high end of this range commonly transport ultramafic xenoliths from mantle depth to the surface. Because mantle-derived xenoliths have not been found at Lathrop Wells, mean ascent rates at mantle depths were probably in the cm/s to 0.1 m/s range. At shallow depth a transition from incompressible to compressible flow occurs due to the exsolution of volatiles triggered by magma decompression. The flow regime changes from a single phase melt flow to one with an increasing larger volume fraction of bubbles. Finally, when the volume fraction of bubbles exceeds the fragmentation limit, the flow becomes significantly compressible and accelerates upon further decompression. The incompressible to compressible flow transition is relevant to the analysis of magmatic hazards for the proposed repository.

10.6.1 Magma Ascent: Incompressible Flow and Thermal Regime

The transport of melt in the incompressible regime has been discussed by many authors including Shaw, 1980; Pollard, 1987; Spence and Turcotte, 1990; Lister, 1990, 1991; Lister and Kerr, 1991; Rubin, 1995; Takada, 1994 and Heimpel and Olson, 1994. Here, we draw on these to present a scale analysis specifically applicable to the eruption at Lathrop Wells. In general, two
source conditions pertinent to magma transport serve as reasonable idealizations: (a) a constant volume source condition in which the magma source is rapidly depleted compared to the time required for magma transport through the crack network and (b) a constant discharge (volumetric rate) condition, applicable provided flow into the magma fracture plexus proceeds long after the fracture has opened. In the case of the relatively small volume (~$10^8$ m$^3$) Lathrop Wells eruption, condition (a) seems more applicable since the volume of a hypothetical single fracture of strike length 1 km and width of several meters extending to the source depth of ~ 50 km is an appreciable fraction of the entire Lathrop Wells eruptive volume. Hence buoyant fractures with constant melt volume that close at the tail are considered appropriate to magma transport feeding the Lathrop Wells eruptive fissure-vent system.

There are two limiting models for the control of the propagation speed of a magma-filled crack. If magma viscosity is the limiting factor, then although the fracture mechanics determines the crack-tip shape, the effect of the crack-tip fracture resistance on the crack propagation velocity is negligible. Models based on this concept give rise to fractures with a slightly bulging head that tapers to a narrow conduit. The crack propagation velocity is equivalent to the Poiseuille velocity of magma in the crack. In this case, the relationship between the average melt ascent velocity, $\bar{v}$, fracture width, $h$, melt viscosity, $\eta$ and driving pressure gradient, $\Delta p/\ell$ is

$$\bar{v} = \frac{h^2 \Delta p}{12 \eta \ell}$$

(1)

If it is assumed that melt buoyancy drives the flow ($\Delta p/\ell\sim \Delta \rho g$), then adopting the scale parameters $h=1$ m, $\eta=500$ Pa s and $\Delta \rho= 200$ kg/m$^3$, a typical crack propagation rate, equivalent to the magma ascent speed for this model is $\bar{v} = 0.3$ m/s. In contrast, for buoyancy-driven crack propagation where the fracture resistance of the solid is important, the fracture velocity, $V_F$ depends on the buoyancy of the melt and the elastic properties of the surrounding lithosphere. In particular, Heimpel and Olson (1994) found that when the melt volume in the buoyancy-driven
fracture is sufficient for $K_I$, the mode I (tensile) crack-tip stress intensity factor \cite{Atkinson,1987} to exceed $K_{IC}$, the mode I critical fracture toughness, unstable dynamic crack-tip propagation results. In typical dynamic fracture applications, the load is applied externally such as by external torsion, and the fracture propagation velocity rapidly approaches the elastic wave velocity of the solid, $V_E$, of order several km/s. However, for constant melt volume buoyancy-driven fracture propagation (the condition relevant to magma transport), the load on the crack tip is not externally defined but instead due to the body force associated with melt buoyancy within the propagating crack of height $\ell$. Therefore, for any volume of melt in the crack, the loading configuration reaches a time-averaged steady state in which the crack-tip propagation velocity is matched by the crack-tail closure velocity. The crack-tip propagation velocity in this case is

$$V_F = c \frac{V_E K_i^2}{2 \ell \sigma y} \left( 1 - \frac{K_i^2}{K_{IC}^2} \right)$$  \hspace{1cm} (2)

where the mode I, stress intensity factor $K_i$ is

$$K_i \sim \frac{4}{\pi} \Delta p \sqrt{\pi \ell/2}$$  \hspace{1cm} (3)

In these expressions, $\sigma y$ represents the extensional yield strength of host rock, $V_E$ is the elastic wave propagation velocity of the lithosphere ($\sim 4$ km/s), $\Delta p$ is the difference between the magma pressure (approximately lithostatic) and the crack-normal principal stress and $c$ is a constant of order unity in SI units. Adopting representative scale parameters for basaltic magma transport through the upper mantle lithosphere ($K_{IC} = 300$ MPa m$^{1/2}$, $V_E = 4$ km/s, $\sigma y = 0.3-1$ GPa, $\ell = 1$ km, $\Delta p = 1-10$ MPa), the fracture propagation speed $V_F \sim 0.1$ m/s. One needs to remember that the stress intensity factor and fracture toughness can vary considerably in geologic media due to heterogeneity, the geometry of the crack and ‘process zone’ effects \cite{Lawn and Wilshaw, 1975; Kostrov and Das, 1988; Atkinson and Meredith, 1987; Rubin, 1995}. It appears that magma fracture propagation speeds in the range cm/s to m/s are appropriate for alkali basalt ascent
through the lithosphere. That is, the two physical models controlling ascent velocity, one limited by melt viscosity and the other by fracture propagation resistance, provide roughly comparable ascent speed estimates. The absence of ultramafic xenoliths in the Lathrop Wells eruptive products is consistent with ascent rates near the lower end of the range, perhaps around 0.01-0.1 m/s. Adopting a mean ascent rate of 0.05 m/s gives a melt travel time of about two weeks from a depth of 60 km. During this period roughly half of the initial volume of magma crystallizes at depth to generate small pods or dike selvages of wehrlite (olivine + clinopyroxene cumulates).

Interestingly, such ultramafic xenoliths are common at other Quaternary alkali basalt localities in the western Great Basin. The eruptive volume of Lathrop Wells is 0.12 km$^3$. Based on phase equilibria modeling, about half of the primary melt volume crystallizes at depth. This implies transport of heat between ascending magma and the surrounding cooler lithosphere. The amount of heat transfer is estimated according to

$$Q_{loss} = \rho_m V_{magma} C_p (T_{liquidus} - T_{solidus}) + \Delta h_{crystallization}$$

where $\rho_m$, $V_{magma}$, $C_p$ and $\Delta h_{cry}$ represent the melt density, total volume of magma generated, isobaric specific heat capacity of melt and specific enthalpy of crystallization, respectively.

Adopting the parameters $\rho_m = 2800$ kg/m$^3$, $V_{magma} = 0.12$ km$^3$, $C_p = 1300$ J/kg K, $\Delta h_{cry} = 375$ kJ/kg and a liquidus to solidus temperature interval of 400 K, the total heat loss is $3 \times 10^{17}$ J or about 125 kJ per kg of primary melt. It is reasonable to ask if this inferred heat loss based on the phase equilibria and fracture mechanical picture is consistent with the implied requirements for heat transport. Recall that in order to match the Mg# of the Lathrop Wells basalt, roughly half of the primary melt generated by partial melting should crystallize within the lithosphere during ascent. Is this inference consistent with elementary heat transfer theory? The model involves a swarm [Shaw, 1980; Takada, 1994] of magma-filled cracks (pods) migrating upwards with concomitant heat loss to surrounding cooler lithosphere. To perform the heat transfer scale analysis, we use an estimate of the volume of a single melt-filled crack ($V_c$) based on fracture mechanics to determine
the total number of magma-filled cracks ($n_c$) needed to match the Lathrop Wells eruptive volume accounting for partial solidification based on equilibria constraints. We then use results based on non-isothermal laminar flow of magma in a propagating crack to estimate the duration of magma flow such that the required heat could be extracted. We then compare the calculated duration to the one estimated from magma ascent fracture mechanics.

The volume of a single melt-filled crack is $V_c = \ell w h$ and hence the total number ($n_c$) of propagating magma-filled cracks associated with the Lathrop Wells eruption is $n_c = V_T / V_c$ where $\ell$ is the crack height, $w$ is the crack length (parallel to strike), $h$ is the crack width and $V_T$ represents the total volume of magma that freezes during ascent (roughly equal to $V_{LW}$ based on phase equilibria calculations). The heat loss ($\dot{Q}$) for flow in a single vertically-propagating crack is

$$\dot{Q} = 4kNu \ell w (T_m - T_w) / h$$

(5)

where $Nu$ is the nondimensional Nusselt number, equal to 7.54 [Bird et al., 1987] for planar crack flow, $k$ is the melt thermal conductivity (0.3 W/m K) and the temperature difference is the mean difference between magma and surrounding lithosphere. The total heat loss ($Q_T$) for an ensemble (swarm) of such propagating melt-filled cracks is

$$Q_T = \frac{V_T}{V_c} \dot{Q} t_{\text{event}}$$

(6)

where $t_{\text{event}}$ represents the duration of the swarm migration based on the phase equilibration requirement that sufficient heat is removed from ascending magma to crystallize about half its starting mass (primary melt). Using typical parameters ($\ell=w=1$ km, $h=2$ m, $\Delta T=500$ K), $t_{\text{event}}$ ~one month is estimated. This agrees with estimates based on fluid and fracture mechanics of ascent from 60 km at mean ascent rate in the range 0.01-0.05 m/s (2 months to two weeks). The number of melt-filled cracks that rise through the lithosphere to feed the Lathrop Wells eruption is ~ 60 in this model.

10.6.2 Magma Ascent: Compressible flow and Explosive Eruption Regime
Magma ascent rates are approximately constant along the lithospheric ascent path because magma behaves approximately as an incompressible fluid before volatile-saturation is attained. However, because the solubility of volatile species (e.g., H$_2$O, CO$_2$, SO$_2$) depends strongly on pressure, rising melt eventually becomes saturated with volatiles at some pressure, $p = p_S$, and a discrete fluid phase forms. Volatile-saturation and continued exsolution and growth of fluid bubbles initiate the regime of bubbly multiphase flow. The pressure (translated to depth with knowledge of the local stress field) at which fluid saturation occurs depends on the composition and abundance of volatiles and the dependence of volatile solubility on temperature, pressure and melt composition. At some pressure less than $p_S$ the volume fraction of fluid in magma may exceed the critical threshold for magma fragmentation. This pressure is identified as the fragmentation pressure, $p = p_F$. In this regime, magma is fluid-dominated volumetrically and blobs of melt (pyroclasts) are carried upwards in a rapidly expanding and accelerating fluid phase. The depth at which fragmentation takes place depends on the same parameters that control the volatile-saturation pressure and additionally on whether exsolved volatiles can leak from the magmatic mixture into the surrounding host rhyolitic tuff. The end-member scenarios are: (1) open system behavior in which volatiles are immediately and completely expelled from rising magma upon exsolution from the melt or (2) closed system behavior in which volatiles exsolved from the melt remain within the magmatic mixture. The dynamics of magma in the compressible regime is clearly relevant to volcanic hazard analysis and is discussed briefly below.

### 10.6.3 Magma Ascent: Thermodynamic Volatile Model

The thermodynamic mixed volatile (H$_2$O-CO$_2$) model of Papale [1999] is used to estimate the composition and abundance of dissolved and exsolved volatiles in melt of Lathrop Wells composition as a function of pressure, temperature and volatile bulk composition. Closed system behavior is assumed; the depth estimate for magma fragmentation is therefore a maximal value. Given the total abundance of volatiles and the bulk system mass ratio of H$_2$O to CO$_2$, the
mass fraction of H₂O and CO₂ in coexisting volatile-saturated melt and fluid can be computed. 

Results exhibiting the properties of melt, fluid and magma are portrayed in Figure 10.9 for basaltic magma of Lathrop Wells composition at 1150 °C with 4 wt. % total volatiles (H₂O + CO₂) and a bulk fluid mass ratio H₂O/CO₂ of 20:1. Figure 10.9a,b shows the magma, fluid and melt density and viscosity as a function of pressure. Although the melt viscosity and density increases upon decreasing pressure, the viscosity (not shown) and density of the magma decreases due to the increasing volume fraction of fluid in magma upon decompression. Especially note the very rapid decrease in magma density as magma rises close to the surface. In figure 10.9c, the variation of mass fraction and volume fraction of fluid in magma is given as a function of pressure. The critical volume fraction at fragmentation, θ_{crit} = 0.7 is marked by the vertical line. This critical rheological limit is attained at a magma pressure ~ 17 MPa for the assumed parameters listed on the Figure. Finally, in Figure 10.9d, the fragmentation pressure is plotted versus the total initial dissolved volatile abundance. The fragmentation pressure varies from 4 MPa to 17 MPa at 1150 °C as total volatile concentrations increase from 1 wt. % to 4 wt. %.

Conversion of the saturation pressure, p_S and the fragmentation pressure, p_F to depth beneath Yucca Mountain requires knowledge of the state of stress with depth. Yucca Mountain lies within a region undergoing active ESE-WNW extension [Zoback, 1989]. The state of stress around Yucca Mountain has been studied using hydraulic fracturing stress measurements, breakout and drilling-induced fractures, earthquake focal mechanisms, and fault-slip orientations [Carr, 1974; Springer et al., 1984; Stock et al., 1985; Rogers et al., 1983; Warren and Smith, 1985; Harmsen and Rogers, 1986; Frizzel and Zoback, 1987; Sandia National Laboratories, 1997]. A review of these data is given in Stock and Healy [1988]. Stress measurements at <1.5 km depth give the ratio of horizontal to vertical stress S_h/S_v ~ 0.35-0.7. If it is assumed that S_v is lithostatic (S_v = ρgz, ρ is the density of tuff, and z is depth), then p_S and p_F can be converted to depth. The dynamic constraint on magma pressure (p_m) for crack propagation is that p_m ≥ σ_3.
Consistent with the state of stress at Yucca Mountain, we assume $S_h=\sigma_3$ and $S_v=\sigma_1$ and that $S_h=2/3 S_v$. Accordingly, the relationship between depth beneath Yucca Mountain and magma pressure is approximately $z = \frac{3p_m}{2g\rho_{\text{eff}}}$. Magma fragmentation pressures of 4 MPa and 17 MPa correspond to depths of ~ 270 m and ~ 1300 m for total volatile contents of 1 wt. % and 4 wt. %, respectively assuming closed system (no leakage of volatile from magma to tuff host rock) behavior.

Finally, we can obtain an estimate of the increase in the magma ascent velocity in the compressible flow regime. The theory is based on a model in which volatile-saturated magma, treated as a compressible, viscous, homogeneous pseudofluid, flows upwards in a vertical conduit of constant cross-sectional area [Spera, 1984]. Magma is assumed to flow through the conduit at constant flux $\dot{M} = \rho_o u_o$. The initial magma density and velocity are taken from the incompressible regime with $\rho_o = 2700$ kg/m$^3$ and $u_o = 1$ m/s. Scaling of the conservation of mass expression then leads to an expression for the increase in the magma velocity $\delta u \sim \frac{\dot{M}}{\rho^2} \delta \rho$ between the depth of volatile saturation and the surface. With $\dot{M} = 2.7\times10^3$ kg/m$^2$ s, $\delta \rho \approx 2000$ kg/m$^3$ and $\rho = 250$ kg/m$^3$ (see Figure 10.9a), $\delta u \sim 90$ m/s. Evidently, magma accelerates rapidly as the surface is approached. These magma exit velocities are consistent with eruption plumes on the order of several kilometers in height. Shallow crustal xenoliths of silicic tuff are common in the cinder cone at Lathrup Wells, consistent with the discharge rates inferred above [Doubik and Hill, 1999].

### 10.6.4 Magma Ascent: Compressible Regime Summary

In summary, the presence of a few percent by mass of dissolved volatiles in basaltic melts will insure attainment of volatile-saturation at depths on the order of a few kilometers and magma fragmentation at depths on the order of one kilometer. If volatiles can leak out of rising magma
by expulsion into surrounding country rock, the fragmentation depth will move upwards to shallower depths. The extent of upward migration of the fragmentation depth depends on the rate of fluid expulsion relative to the rate of volatile exsolation. To first order, the rate of fluid exsolation depends on the ascent rate of rising magma. This can be modeled with knowledge of the permeability of the country rock and information on the physical and thermodynamic state of rising magma; this calculation is not performed here. We have shown in section 10.4.1 that dissolved water contents consistent with those inferred for Lathrop Wells are easily achieved provided a small amount of H$_2$O (e.g., 0.2 wt. %) is present in the source peridotite. The ultimate source of these fluids may be related to subduction along the western margin of the North American plate during the Mesozoic and Early Cenozoic although this is speculative. The saturation pressure ($p_S$) and fragmentation pressure ($p_F$) can be converted the depth once the state of stress in a region is specified. Beneath Yucca Mountain, the minimum principal stress is horizontal and the measured state of stress implies that magma fragmentation takes place at depths in the range 0.5 to 1.5 kilometers depending on volatile contents assuming closed system behavior. If volatiles do indeed escape from rising magma, then the depth at which magma fragmentation occurs will be shallower. An eruption with a conduit anywhere within the YMR footprint would be expected to mobilize some fraction of radioactive waste; that fraction is beyond the scope of this study to estimate although we can say that the larger the eruptive volume and the higher the initial volatile-content of the magma the greater amount of radioactive materials would be dispersed. Eruption plume heights scale with the one-fourth root of the mass discharge $\dot{M}A_c$ where $A_c$ is the cross-sectional area of the volcanic conduit, typically in the range $10^2$-$10^4$ m$^2$ and $\dot{M}$ is the mass flux defined in section 10.6.3. Once in the atmosphere, ash dispersal depends very strongly on the structure of winds aloft, the particle size distribution of entrained particles and other parameters.
10.7 Concluding Remarks

In this study we have shown that it is possible to obtain a sketch of the thermodynamics and mechanics of magma generation, partial crystallization, ascent and eruption for small-volume alkali basaltic volcanoes. These volcanoes are typically present in continental extensional provinces such as the Great Basin of western North America where Yucca Mountain lies. We find that melting of a fertile peridotitic source containing a small amount of H$_2$O at depths in the range around 50-60 km followed by significant fractional crystallization (about 50 % by mass) during upward ascent is consistent with the petrology, geochemistry and inferred eruptive history of Lathrop Wells volcano, a 78 ka volcanic construct of tephra and lava about 20 km south of the proposed nuclear waste repository. Ascent rates at depths greater than a few kilometers are in the range 0.01-0.1 m/s and are limited either by the fracture toughness of the surrounding mantle or by the shear viscosity of the melt. Ascent rates are roughly constant as magma rises through the upper mantle and lower crust. At shallow depths (a few kilometers) magma contains about 3 wt. % H$_2$O and becomes volatile-saturated. As magma continues to ascend and decompress the volume fraction of exsolved supercritical fluid phase, dominated by H$_2$O component but also undoubtedly containing small amounts of CO$_2$ and other constituents (e.g., SO$_2$, H$_2$S, CO), rapidly increases approaching and then exceeding the fragmentation limit provided fluid does not leak out of magma into the surrounding volcanic tuff. Around this critical rheological transition from melt-dominated to fluid-dominated magma limit, the density and viscosity of magma rapidly declines. This leads to significant magma acceleration from initial velocity around 1 m/s to final velocity (at the vent) of ~100-200 m/s. For total magma volatile contents around 3 wt. %, the fragmentation limit is reached at about 1 km depth, well below the depth of the proposed repository at Yucca Mountain of 300 m depth. If volatiles can escape rising magma, fragmentation depths move upwards to shallower depths.

An analytical survey of the consequences of repository disruption by magmatic process has been presented elsewhere [Detournay et al., 2003] although many important issues remain to
be more fully explored quantitatively. Although such an analysis is beyond the scope of the present study, it is clear that quantitative conceptual models of alkali basalt petrogenesis of the type outlined in this study form the basis of informed volcanic hazard analysis.

FURTHER READING


REFERENCES CITED


Table 1
Anhydrous composition of peridotites used in the partial melting simulations. IB-33 is a depleted peridotite from the Lunar Crater Volcanic Field, Nevada, U.S.A., PA-12 is a moderately fertile lherzolite from Peridot Mesa, San Carlos Volcanic Field, Arizona, U.S.A and KLB-1 a fertile peridotite from Kilbourne Hole, New Mexico, U.S.A. These peridotite samples are found as xenoliths in Quaternary alkali basalts in the Great Basin of western North America.

<table>
<thead>
<tr>
<th>Wt %</th>
<th>IB-33</th>
<th>PA-12</th>
<th>KLB-1</th>
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<td>SiO₂</td>
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Table 2
Mineralogy and fertility measures for peridotite starting compositions used in the partial melting simulations. Fertility increases from 1B-33 (depleted) to PA-12 (moderately fertile) to KLB-1 (fertile). To each anhydrous bulk composition in Table 1, 0.2 wt % H$_2$O has been added. At 1.5 GPa amphibole (amp) and phlogopite (phl) are present in the fertile compositions at the solidus. Garnet (gt) and phlogopitic mica are present at 3.5 GPa.

<table>
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<tr>
<th>Sample</th>
<th>Mass fraction$^2$</th>
<th>Wt. % alkalies</th>
<th>Wt. ratio FeO/MgO</th>
<th>Wt. % MgO</th>
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<td></td>
<td>1.5 GPa</td>
<td>3.5 GPa</td>
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<td></td>
</tr>
<tr>
<td>1B-33 Lunar Crater, Nevada, USA</td>
<td>0.036 ol,opx, gt, cpx</td>
<td>0.067 ol, opx, gt, cpx, phl</td>
<td>0.08</td>
<td>0.16</td>
</tr>
<tr>
<td>PA-12 San Carlos, Arizona, USA</td>
<td>0.083 ol, opx, cpx, amp, sp, phl</td>
<td>0.152 ol, opx, gt, cpx phl, sp</td>
<td>0.43</td>
<td>0.19</td>
</tr>
<tr>
<td>KLB-1 Kilbourne Hole, New Mexico, USA</td>
<td>0.190 ol, opx, cpx, amp, sp, phl</td>
<td>0.325 ol, opx, cpx, gt, sp, phl</td>
<td>0.33</td>
<td>0.21</td>
</tr>
</tbody>
</table>

$^1$ Order of phase listing is from most to least modal abundance at the prescribed pressure
$^2$ Phase abundances are at solidus temperature at indicated pressure with 0.2 wt % H$_2$O added to the anhydrous compositions of Table 1. The redox condition is QFM-1 in all cases.
Table 3
Initial trace element concentrations used for trace element modeling of combined batch partial melting of KLB-1 fertile peridotite and subsequent fractional crystallization. Results presented in Figure 8. Trace element concentrations from McDonough, 1990.

<table>
<thead>
<tr>
<th>Element</th>
<th>Concentration (ppm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rb</td>
<td>1.90</td>
</tr>
<tr>
<td>Ba</td>
<td>33</td>
</tr>
<tr>
<td>Th</td>
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</tr>
<tr>
<td>Nb</td>
<td>4.8</td>
</tr>
<tr>
<td>Ta</td>
<td>0.4</td>
</tr>
<tr>
<td>La</td>
<td>2.60</td>
</tr>
<tr>
<td>Ce</td>
<td>6.29</td>
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<tr>
<td>Sr</td>
<td>49</td>
</tr>
<tr>
<td>Nd</td>
<td>2.67</td>
</tr>
<tr>
<td>Sm</td>
<td>0.47</td>
</tr>
<tr>
<td>Zr</td>
<td>21</td>
</tr>
<tr>
<td>Hf</td>
<td>0.27</td>
</tr>
<tr>
<td>Eu</td>
<td>0.16</td>
</tr>
<tr>
<td>Tb</td>
<td>0.07</td>
</tr>
<tr>
<td>Y</td>
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</tr>
<tr>
<td>Yb</td>
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</tr>
<tr>
<td>Lu</td>
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</tr>
<tr>
<td>Co</td>
<td>112</td>
</tr>
<tr>
<td>Cr</td>
<td>2690</td>
</tr>
<tr>
<td>Ni</td>
<td>2160</td>
</tr>
<tr>
<td>Sc</td>
<td>12.2</td>
</tr>
<tr>
<td>V</td>
<td>56</td>
</tr>
<tr>
<td>Zn</td>
<td>65</td>
</tr>
</tbody>
</table>
Table 4
Solid-melt partition coefficients (K_{sm}) used in partial melting and fractional crystallization trace element modeling results presented in Figure 8. Data sources: GERM website (http://www.earthref.org/GERM/), Henderson, 1982, Best and Christiansen, 2001.
Phl (trioctahedral phyllosilicate), Cpx (clinopyroxene), Amp (amphibole), Ol (olivine), Opx (orthopyroxene), Sp (spinel).

<table>
<thead>
<tr>
<th>Element</th>
<th>Phl</th>
<th>Cpx</th>
<th>Amp</th>
<th>Ol</th>
<th>Opx</th>
<th>Sp</th>
</tr>
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<tbody>
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<td>0.0002</td>
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</tr>
<tr>
<td>Ba</td>
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<td>0.00011</td>
<td>0.1</td>
<td>0.0000022</td>
<td>0.0000036</td>
<td>0.028</td>
</tr>
<tr>
<td>Th</td>
<td>0.12</td>
<td>0.04</td>
<td>0.11</td>
<td>0.03</td>
<td>0.013</td>
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<tr>
<td>Nb</td>
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<td>0.8</td>
<td>0.11</td>
<td>0.15</td>
<td>0.4</td>
</tr>
<tr>
<td>Ta</td>
<td>0.56</td>
<td>0.261</td>
<td>0.62</td>
<td>0.17</td>
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<tr>
<td>La</td>
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<td>0.002</td>
<td>0.045</td>
<td>0.0004</td>
<td>0.0003</td>
<td>0.0029</td>
</tr>
<tr>
<td>Ce</td>
<td>0.03</td>
<td>0.017</td>
<td>0.09</td>
<td>0.0001</td>
<td>0.0008</td>
<td>0.01</td>
</tr>
<tr>
<td>Sr</td>
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<td>0.077</td>
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<td>0.0003</td>
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<td>0.01</td>
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<tr>
<td>Sm</td>
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<td>0.09</td>
<td>0.033</td>
<td>0.00014</td>
<td>0.0023</td>
<td>0.0072</td>
</tr>
<tr>
<td>Zr</td>
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<td>0.02</td>
<td>0.02</td>
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<tr>
<td>Hf</td>
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<td>0.19</td>
<td>0.001</td>
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<tr>
<td>Eu</td>
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<td>0.09</td>
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<td>0.001</td>
<td>0.0033</td>
<td>0.01</td>
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<tr>
<td>Tb</td>
<td>0.7</td>
<td>0.28</td>
<td>0.32</td>
<td>0.0015</td>
<td>0.019</td>
<td>0.01</td>
</tr>
<tr>
<td>Y</td>
<td>0.018</td>
<td>0.467</td>
<td>0.4</td>
<td>0.009</td>
<td>0.18</td>
<td>0.0039</td>
</tr>
<tr>
<td>Yb</td>
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<td>0.48</td>
<td>0.46</td>
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<tr>
<td>Lu</td>
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<td>0.4</td>
<td>0.018</td>
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<tr>
<td>Co</td>
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<td>Cr</td>
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<td>30</td>
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<td>11</td>
</tr>
<tr>
<td>Ni</td>
<td>1.3</td>
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<td>1</td>
<td>6</td>
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<tr>
<td>Sc</td>
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<td>0.67</td>
</tr>
<tr>
<td>V</td>
<td>0.5</td>
<td>1.81</td>
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<td>0.06</td>
<td>0.6</td>
<td>0.5</td>
</tr>
<tr>
<td>Zn</td>
<td>7</td>
<td>0.49</td>
<td>0.69</td>
<td>0.86</td>
<td>0.41</td>
<td>2.6</td>
</tr>
</tbody>
</table>
Figure 10.2

Trachybasalt
Basanite
Picro-basalt
Basalt
Basaltic trachyandesite

Batch melting (fertile, 1.5 GPa)
Fractional melting (fertile, 1.5 GPa)

5 % melt
10 % melt
15 % melt
Figure 10.3
Figure 10.4
(a) Batch melting
P: 1.5 GPa
Initial H₂O: 0.2 wt. %
$\text{f}_0\text{O}_2$ buffer: QFM-1

(b) Fractional crystallization
P: 1.0 GPa
Initial H₂O: 1.74 wt. %
$\text{f}_0\text{O}_2$ buffer: QFM-1

Figure 10.6
Figure 10.7
Figure 10.8

- Mg # = 52; $f_m = 0.55$
- Liquidus temperature; $f_m = 1$
- Lathrop Wells data
Figure 10.9
Figure captions

Fig. 1. Shaded relief map of Pliocene and Quaternary basaltic volcanoes in the Yucca Mountain region including buried volcanic rocks identified by geomagnetic methods and sampled by drilling. The proposed Yucca Mountain repository footprint is in yellow. Purple dashed lines are Miocene caldera boundaries. Miocene volcanoes are shown in yellow and green. Pliocene volcanoes are shown in pink. Quaternary volcanoes appear in red. 1.1 Ma episode: Makani volcano (MC), Black Cone (BC), Red Cone (RC), and NE and SW Little Cones (LC). 0.35 Ma episode: Little Black Peak (LBP), Hidden Cone (HC). 77 ka Lathrop Wells volcano (LW). Blue lines are major faults bounding the Armagosa Trough. Map provided by Dr. Frank V. Perry of LANL. Efforts of the GIS group at LLNL are gratefully acknowledged.

Fig. 2. Comparison of isobaric fractional (grey dashed line) and batch (solid black line) partial melting phase equilibria calculations using the pMELTS algorithm for fertile peridotite bulk composition KLB-1 (Hirose and Kushiro, 1993). Oxygen fugacity is fixed along the QFM-1 buffer henceforth. Increasing wt. % MgO and decreasing wt. % (Na$_2$O + K$_2$O) correspond to increasing magma temperature. The beginning of each pMELTS trend is labelled $T_S$. Cross marks within square outlines signify 5, 10, and 15 % melting, with 5 % melting closest to the solidus. The termination of each pMELTS trend corresponds to a melt fraction of 20 %. Shaded region represents field of Lathrop Wells data (Perry and Straub, 1996; Valentine and Perry, 2007). The data field is not expected to coincide with the pMELTS trends due to the non-primary nature of Lathrop Wells basalt (see text for further details). (a) Total alkalies-silica diagram using classification scheme of Le Maitre et al. (1989). (b) MgO vs. FeO. (c) MgO vs. Al$_2$O$_3$. (d) MgO vs. SiO$_2$. (e) MgO vs. CaO.

Fig. 3. Results of isobaric pMELTS batch partial melting calculations at 1.5 (large symbols) and 3.5 (small symbols) GPa for depleted, moderately fertile and fertile peridotite. Compositions are given in Table 2. Dot-dashed line: depleted peridotite 1B-33 (Bergman, 1982); dashed grey line: moderately fertile peridotite PA-12 (Frey and Prinz, 1978); solid black line: fertile peridotite KLB-1 (Hirose and Kushiro, 1993). All other parameters and abbreviations are identical to Fig. 2. (a) SiO$_2$ vs. total alkalies. (b) MgO vs. FeO. (c) MgO vs. Al$_2$O$_3$. (d) MgO vs. SiO$_2$. (e) MgO vs. CaO.

Fig. 4. pMELTS isobaric batch partial melting trends based on fertile peridotite KLB-1 (Hirose and Kushiro, 1993) for varying pressure. Solid black line: 1.5 GPa; solid grey line: 2.0 GPa; dotted black line: 2.5 GPa; dot-dashed line: 3.0 GPa; dashed grey line: 3.5 GPa. All other parameters and abbreviations are identical to Fig. 2. (a) SiO$_2$ vs. total alkalies. (b) MgO vs. FeO. (c) MgO vs. Al$_2$O$_3$. (d) MgO vs. SiO$_2$. (e) MgO vs. CaO.

Fig. 5. Results of numerical experiments designed to examine batch partial melting pMELTS liquid compositions with varying initial bulk water contents. The source composition is fertile peridotite KLB-1 (Hirose and Kushiro, 1993). Dashed black line: anhydrous; dotted grey line: 0.1 wt. % H$_2$O; solid black line: 0.2 wt. % H$_2$O; dashed grey line: 0.6 wt. % H$_2$O. All other parameters and abbreviations are identical to Fig. 2. (a) SiO$_2$ vs. total alkalies. (b) MgO vs. FeO. (c) MgO vs. Al$_2$O$_3$. (d) MgO vs. SiO$_2$. (e) MgO vs. CaO. Hereafter, the 0.2 wt. % H$_2$O case is used as the reference case.

Fig. 6. Phase proportions as a function of temperature (T) based on (a) batch partial melting of fertile peridotite KLB-1 (Hirose and Kushiro, 1993) plus 0.2 wt. % H$_2$O at 1.5 GPa. The 10 % partial melt has a dissolved H$_2$O content of 1.74 % at about 1310 °C at 1.5 GPa and is used as the starting composition for the fractional crystallization simulation (b) Mineral proportion diagram for isobaric (1 GPa) fractional crystallization starting with liquid generated at 10 % batch melting (Figure 6a). Crystallizing phases are olivine, clinopyroxene, and spinel, in the order given. Fractionation is stopped when the Mg# equals 52, the observed Mg# for Lathrop Wells basalt. Abbreviations: Amp, amphibole; Cpx, clinopyroxene; Ol, olivine; Opx, orthopyroxene; Phl, phlogopite; Sp, spinel.
Fig. 7. MgO versus oxide (wt. %) variation diagrams showing Lathrop Wells data (shaded field; Perry and Straub, 1996) and (p)MELTS calculation results. Predicted trends represent evolution of a 10 % KLB-1 (Hirose and Kushiro, 1993) primary melt formed by batch partial melting at 1.5 (dashed black line) and 3.5 (dotted grey line) GPa by isobaric, closed-system fractional crystallization (FC) at 1.0 GPa (solid grey line) and 0.5 GPa (solid black line). Thinner lines show the results of partial melting from Figure 4. A cross marks the beginning of each fractional crystallization trend. Decreasing wt. % MgO and increasing wt. % (Na\textsubscript{2}O + K\textsubscript{2}O) correspond to decreasing magma temperature. The termination of each fractional crystallization trend is labelled (Mg\# = 52). See text for details. (a) SiO\textsubscript{2} vs. total alkalies. (b) MgO vs. FeO. (c) MgO vs. Al\textsubscript{2}O\textsubscript{3}. (d) MgO vs. SiO\textsubscript{2}. (e) MgO vs. CaO. Note that for all oxides except alumina, the termination of the computed fractionation path lies in or very near the region of the Lathrop Wells composition for fractionation at 1 GPa of the 10% partial melt of fertile peridotite generated at 1.5 GPa.

Fig. 8. Comparison of Lathrop Wells trace element data (absolute abundances; Perry and Straub, 1996) and calculation results based on 10 % isobaric batch partial melt of KLB-1 (Hirose and Kushiro, 1993) followed by closed-system fractional crystallization isobarically at 1 GPa terminating at a liquid Mg\# = 52. Symbol T\textsubscript{l} gives the concentration of the trace element at the beginning of isobaric fractionation. The large X gives the trace element concentration at the end of crystal fractionation when the Mg\# = 52, the low end of the small range (52-56) for Lathrop Wells. See text for details.

Fig. 9. Variation of magma properties for Lathrop Wells basaltic magma as function of pressure. (a) melt viscosity and magma density vs. pressure. (b) melt density and fluid (vapor) density vs. pressure. (c) mass fraction and volume fraction of exsolved supercritical fluid vs. pressure. (d) fragmentation pressure of Lathrop Wells basalt magma as a function of total volatile content. The fragmentation pressure is defined as the pressure at which the volume fraction of fluid equals 0.7. See text for details.