Geologic Setting of the Snake River Plain Aquifer and Vadose Zone

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ABSTRACT

The Snake River Plain aquifer owes its existence and abundance to a unique sequence of tectonic, volcanic, and sedimentologic processes associated with the migration of the North American tectonic plate southwestward across the Yellowstone hotspot, or mantle plume. The basalt lava flows that host the aquifer and comprise the overlying vadose zone are very porous and permeable due to emplacement processes and fracturing during cooling. Rubble zones between lava flows and cooling fractures allow very rapid flow of water in the saturated zone, rapid infiltration of water and contaminants, and deep penetration of air into the vadose zone. Alluvial, eolian, and lacustrine sediments interbedded within the basalt sequence are generally fine-grained, commonly serving as aquitards below the water table, and affecting infiltration and contaminant transport in the vadose zone. The subsiding eastern Snake River Plain (ESRP) and the high elevations of the surrounding recharge areas comprise a large drainage basin that receives enormous amounts of precipitation and feeds high-quality groundwater into the aquifer. Northeast–southwest directed extension of the ESRP produces significant anisotropy to the hydraulic conductivity of the rocks. High heat flow and upwelling of geothermal fluids from the crust beneath the aquifer affect geothermal gradients, aquifer temperatures, and solute chemistry.

The eastern Snake River Plain, located in southeastern Idaho (Fig. 1), is an elongate, down-warped basin containing a 1-km-thick sequence of Quaternary and late Tertiary basalt lava flows overlying late Tertiary rhyolitic volcanic rocks. It extends across southern Idaho from the Yellowstone Plateau to the Twin Falls area. The Idaho National Engineering and Environmental Laboratory (INEEL), a 2300-km^2 reservation run by USDOE, is located near the northwestern boundary of the ESRP (Fig. 1). Throughout most if its history, the INEEL has been used for the testing of nuclear reactors, and in recent years, the disposal and remediation of various types of nuclear and chemical wastes have become important issues. Much of the vadose zone and aquifer research conducted at the site is focused on increasing the understanding of water and contaminant migration as it relates to these disposal and remediation issues.

The vadose zone at the INEEL ranges in thickness from 60 to >300 m and exhibits extreme spatial heterogeneity in texture, composition, physical properties, and structure. The underlying Snake River Plain aquifer, a world-class aquifer in terms of water volume and discharge, flows southwestwardly from recharge areas on the Yellowstone Plateau to discharge areas in the Snake River canyon near Twin Falls, ID. The aquifer is up to 400 m thick, underlies 26 000 km^2, and contains about 100×10^9 m^3 (100 million acre-feet) of water. Transit time for water from recharge to discharge, a distance of about 320 km, is about 300 yr. Many of the problems associated with characterization and remediation in this complex environment are discussed in other papers in the special section of this issue. The purpose of this paper is to provide the regional geologic setting, origin, and evolution of the Snake River Plain aquifer and vadose zone. It summarizes the current understanding of the processes that produced the ESRP, its world-class aquifer, and its complex and challenging vadose zone.

BRIEF HISTORY OF GEOLOGIC AND HYDROLOGIC INVESTIGATIONS

The Snake River Plain has a long history of geologic and hydrologic investigations, beginning about a century ago when Israel Russell (1902) described the basic geology and hydrology. Russell, and Kirkham (1931), recognized that downwarping of the crust produced the low elevation, low relief basin surrounded by mountain ranges of the Northern Rocky Mountains. This downwarping characteristic sets the ESRP apart from continental rift valleys, such as the East African Rift, and it was not until the 1970s and 1980s that the true nature and origin of the downwarp was recognized. Between 1900 and 1970 geologic studies focused mostly on volcanism in the Craters of the Moon and Mud Lake areas (Fig. 2) (Stearns, 1924, 1926, 1928, 1963; Murtaugh, 1961) and petrogenesis of basalt lava flows (Powers, 1960a, 1960b; Stone, 1967). A few studies dealt with hydrology of the ESRP (Stearns, 1936; Stearns et al., 1938, 1939; Nace, 1958, Olmsted, 1962). In the 1960s geophysical and tectonic characteristics of the plain started receiving attention (LaFehr and Pakiser, 1962), and a summary of the Quaternary sedimentation and the stratigraphy of Quaternary basalt lava flows was published by Malde (1965).

In the 1970s and 1980s, researchers began to focus in earnest on the origin and tectonic evolution of the ESRP, on the origin and age of magmas, and on the volcanic processes that produced the landscapes. Much of the work used petrology and age of volcanic rocks to make inferences about the origin of volcanism of the plain (Leeman, 1974, 1982a, 1982b, 1982c, 1982d; Stout and Nicholls, 1977; Stout et al., 1994; Kuntz et al., 1986a, 1986b), and some used the volcanic processes on the plain as analogy for planetary volcanic processes (Greeley and King, 1977; Greeley, 1982). However, a big leap forward in understanding the tectonic evolution of the ESRP came from geophysics (Sparlin et al., 1982; Ma-

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Abbreviations: ESRP, eastern Snake River Plain; INEEL, Idaho National Engineering and Environmental Laboratory.

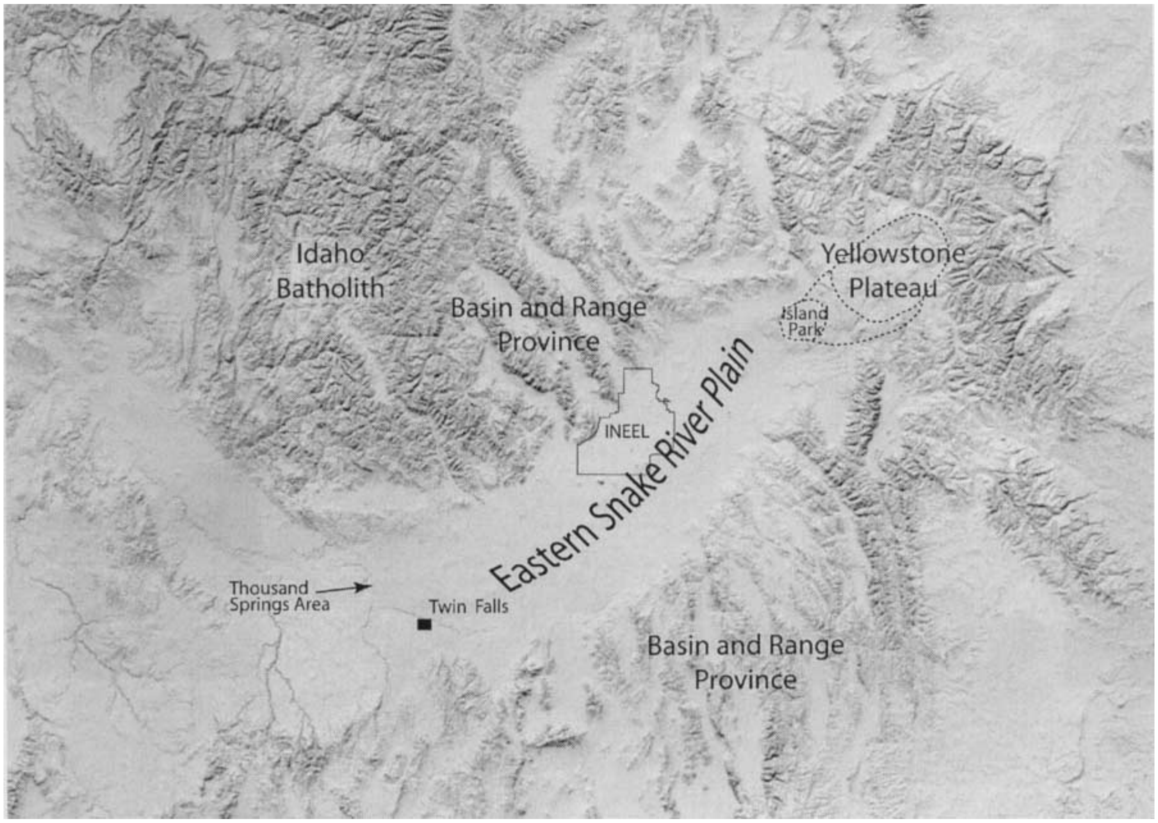

Fig. 1. Shaded-relief topographic map of the eastern Snake River Plain and adjacent mountain ranges of the northern Basin and Range province and the Northern Rocky Mountains.

bey, 1982; Brott et al., 1981; Blackwell, 1989), regional age trends in silicic volcanic rocks (Armstrong et al., 1975; Suppe et al., 1975), and inferences about convection plumes, or hotspots, in the earth's mantle (Morgan, 1972a, 1972b). The understanding of mantle and crustal processes leading to the formation of the ESRP that was gained from these investigations was further developed in the late 1980s and 1990s. The seismic activity on faults adjacent to the Plain were related to the migration of the Yellowstone hotspot (Anders et al., 1989; Anders and Sleep, 1992; Pierce and Morgan, 1992) and to magmatic processes on the plain (Rodgers et al., 1990; Smith et al., 1996; Parsons et al., 1998). The mechanism of subsidence of the Plain during and after the passage

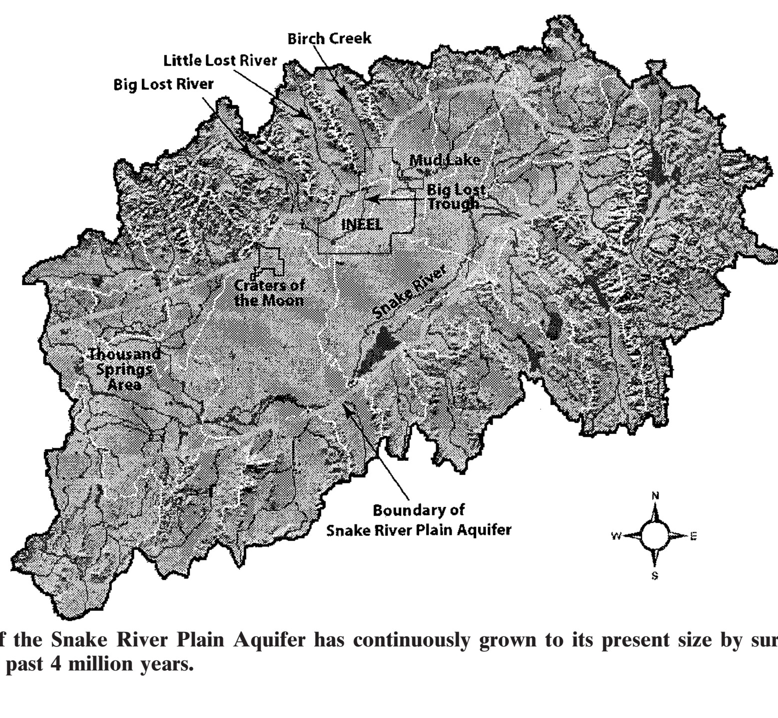

Fig. 2. The drainage basin of the Snake River Plain Aquifer has continuously grown to its present size by surface subsidence of the eastern Snake River Plain for the past 4 million years.

of the hotspot through the ESRP were investigated by McQuarrie and Rodgers (1998). In addition, several authors critically examined the ideas of mantle hotspot origin of the ESRP (Smith and Braile, 1994; Humphreys et al., 2000; Christiansen et al., 2002). Also, the understanding of the mechanisms of eruption and lava flow processes were improved and broadened (Kuntz, 1992; Kuntz et al., 1992; Malde, 1991).

In parallel with these tectonic and volcanic investigations, the Snake River Plain aquifer received increasing study (Nace, 1958; Nace et al., 1975; Whitehead, 1992; Lindholm et al., 1983; Lindholm and Goodell, 1986). Also, several workers began to grapple with modeling of groundwater flow and contaminant movement in the Snake River Plain aquifer (Garabedian, 1992; Robertson, 1974, 1977; Gego et al., 2002), and others used distribution of solutes and isotopes in aquifer waters to infer flow directions and flow velocities (Johnson et al., 2000; Roback et al., 2001; Busenberg et al., 2001; McLing et al., 2002).

In addition, many investigations have focused on correlation of basalt lava flows in the INEEL area. These studies have used paleomagnetism (Champion et al., 1988), age and petrography (Lanphere et al., 1993, 1994), physical volcanology (Wetmore, 1998; Wetmore et al., 1997), and chemical composition (Reed et al., 1997; Hughes et al., 2002) of lava flows to establish lateral correlation. Scarcity of core samples from wells (and thus materials for determination of magnetic, chemical, and petrographic characteristics) limits the scope of these studies. In an effort to overcome this limitation, several studies have focused on the use of borehole γ logs, an almost universal data set for wells at INEEL, to establish correlations (Anderson, 1991; Anderson and Lewis; 1989; Anderson and Bowers, 1995). The most exhaustive summary of basalt stratigraphy based on γ logs appears in Anderson et al. (1996). Many stratigraphic cross sections showing correlations across parts of the INEEL are presented in these publications.

GENESIS AND EVOLUTION OF THE EASTERN SNAKE RIVER PLAIN— A SHORT SUMMARY

The ESRP owes its existence to Tertiary and Quaternary tectonic, volcanic, and sedimentary processes that profoundly affected the crust during and after the movement of the North American tectonic plate southwestwardly across the Yellowstone mantle plume or hotspot (Leeman, 1982a; Smith and Braile, 1994). The hotspot now resides beneath the Yellowstone Plateau and is responsible for the young silicic volcanism and abundant geothermal features present there (Smith and Braile, 1994). Silicic volcanic rocks (rhyolites) analogous to those at the surface of the Yellowstone Plateau today underlie the basalt lavas of the ESRP (Morgan et al., 1984; Hackett and Morgan, 1988; Hackett and Smith, 1992), and record an early stage in the formation of the Plain. This early silicic stage of volcanism occurred 4 to 10 million years ago when the Yellowstone hotspot was beneath the ESRP. There is strong evidence that silicic volcanic rocks are distributed in a temporal and spatial transgressive sequence beginning about 17 million years ago in north-central Nevada and ending at the Yellowstone Plateau today (Armstrong et al., 1975). During the time that the hotspot was beneath the INEEL area of the Plain, the area may have been a high plateau like Yellowstone today, and large, explosive volcanic eruptions produced thick deposits of rhyolitic ash flow tuffs and lava flows (Hackett and Morgan, 1988). During this time profound modifications of the middle crust of the ESRP occurred. Enormous volumes of mantle-derived basaltic magma moved upward through the lower crust and stalled in the middle crust at depths of 8 to 20 km below the surface (Leeman, 1982a; Sparlin et al., 1982; Smith and Braile, 1994). It cooled and solidified there, and interacted with crustal rocks to produce large volumes of silicic magma that rose into large magma chambers near the surface. The breaching of the tops of these large magma chambers and the floundering of roof blocks into the chambers initiated large explosive eruptions of rhyolitic tuffs and lavas, and produced calderas analogous to those at Yellowstone. During residence above the hotspot and after it moved off the hotspot about 4 million years ago, the shallow magma chambers and the mid-crustal basaltic (gabbroic) magma chambers solidified and cooled, and the ESRP began a 4 million-year period of crustal subsidence, surface subsidence, and basaltic volcanism. That subsidence continues today and is driven by isostatic adjustments of the crust to the dense mid-crustal gabbroic mass (McQuarrie and Rodgers, 1998) and by contraction as the crustal column cools (Brott et al., 1981). Because the upper mantle beneath the ESRP retains relatively high temperatures, partial melting of mantle material continues to produce basaltic magmas that rise through the crust to erupt as lavas that fill the subsiding basin. The subsidence and basalt volcanism take place within the extensional regime of the northern Basin and Range province (Smith et al., 1996), and the ESRP is being continuously stretched in a northeast–southwest direction, parallel to its length. Within the ESRP that extension is accommodated by intrusion of northwest-trending basalt dikes during volcanic episodes, whereas outside the Plain the extension is accommodated by north to northwest-trending normal faulting (Parsons et al., 1998).

To summarize, the INEEL area of the ESRP has been profoundly affected by its ride over the Yellowstone hotspot during the past few million years. First, the middle crust was intruded and partially melted by mantle-derived basaltic magmas. This lead to emplacement of large silicic magma chambers at very shallow depth and to violent caldera-forming rhyolitic eruptions. Then, as the area moved southwestwardly off of the hotspot, subsidence of the surface produced the elongate basin of the ESRP and persistent basaltic volcanism filled the basin with a thickness of up to 1 km or more of basalt lavas. At the same time, erosion of the surrounding mountains provided alluvial, lacustrine, and eolian clastic sediments to the basin to form sediment interbeds within the thick basaltic lava flow sequence. These concepts have many details, corollaries, and implications that will not be discussed here, but are eloquently described elsewhere (Armstrong et al., 1975; Leeman, 1982a, 1982b, 1982c, 1982d; Anders et al., 1989; Anders and Sleep, 1992; Pierce and Morgan, 1992; Hackett and Smith, 1992, Smith and Braile, 1994; Humphreys et al., 2000; Christiansen et al., 2002). The remainder of this paper will focus on those aspects of the origin and evolution of the ESRP that have direct relevance or consequence to the aquifer and vadose zone in the INEEL area.

CONSEQUENCES OF THE STYLE OF BASALT VOLCANISM ON THE ESRP

The upper mantle and lower crust beneath the ESRP are hotter than elsewhere in the Basin and Range province and Northern Rocky mountains (Brott et al., 1981; Blackwell, 1989) and continue to produce small volumes of basaltic magma. Because of the continual extension experienced by the ESRP, magma bodies move upwards toward the surface while they are still small, and the overall result is numerous, small-volume eruptions instead of a few big ones. This is an important distinction between the basalt sequence on the ESRP and other large basalt provinces in the world (e.g., Columbia Plateau, Deccan Traps). As a result, the Snake River Plain aquifer and its vadose zone are hosted by a sequence of thousands of basalt lava fields, each with numerous lava flows a few meters thick (Table 1). The largest lava fields exposed at the surface contain about 6 km^3 of basaltic volcanic rocks and cover a few hundreds of square kilometers, with lava flows that travel up to about 35 km from their vents (Kuntz et al., 1992, 1994).

The style of eruption of these small-volume lava fields contributes to a very porous and permeable host rock for the Snake River Plain aquifer, and for the overlying vadose zone. Individual lava flows tend to exhibit a characteristic distribution of "facies" (Fig. 3). Each flow has a rubble zone at its base where solidified blocks that fall from the flow front are overridden by the advancing flow. These rubble zones commonly are a meter or so thick, contain decimeter-scale blocks with large cavities between them, and can have up to 50% porosity. The lava flows themselves develop a characteristic fracture pattern as they crystallize and cool, with an upper and lower zone of platy fractures (parallel to upper and lower surfaces) and close-spaced (a few inches) columnar joints (perpendicular to upper and lower surfaces). Inside, the central, massive part of the flow develops a system of coarse (usually nearly vertical, meter-scale) columnar joints. In addition, the near-vent areas of the flows contain very porous and permeable layers of cinders and ash, as well as thin sheets (inch to foot-scale) of lava with subhorizontal open fractures between them (called shelly pahoehoe). Also, the lava flows are commonly vesicular with upper and lower zones rich in gas bubbles. Since the fracturing develops as a result of stresses imparted during cooling, each flow has its own distribution of fracture spacings and attitudes that bear little relationship to those of the overlying and underlying flows. The largest of the ESRP lava flows are fed by lava tubes that carry lava from the vent area to the flow front. The lava tubes commonly are several meters in diameter and, on cessation of the eruption, drain out to the flow front, leaving a hollow tube encased within the lava flow.

Table 1. Idaho National Engineering and Environmental Laboratory area lava flow dimensions (from Hackett et al., 2004).

	Length	Area	Thickness
	km	km^2	m
Minimum	0.1	0.5	1
Maximum	31	400	34
Mean	12.4	96.5	–
Median	10	70	7
Standard deviation	7.9	94.2	–
Number of measurements	46	43	641

Because upper mantle temperatures are greatest beneath the northeast-trending axis of the ESRP (Humphreys and Dueker, 1994), most volcanic activity occurs in the so-called Axial Volcanic Zone and in those portions of the volcanic rift zones closest to the axis of the ESRP (Fig. 4). The greatest concentration of volcanic vents occurs along this zone and in the nearby portions of the volcanic rift zones, and shorter recurrence intervals for volcanism occur there than elsewhere on the Plain (Hackett et al., 2004). Consequently, a northeast-trending axial ridge and several northwest-trending volcanic rift zones made up of a high percentage of vent-facies pyroclastic rocks and shelly pahoehoe sustain the highest elevations of the Plain (Fig. 3 and 4). These facies are especially porous and permeable and provide zones of high permeability for both saturated and unsaturated migration of water and contaminants. Because of the high elevation maintained along the axis of the ESRP, it collects much less sediment than other areas of the Plain, and only sparse, thin eolian deposits occur within the basalt sequence. The paucity of interbeds also contributes to high permeabilities in this part of the Plain. The axial ridge also imparts an anisotropy to the shapes of lava flows that originate there and to the orientations of associated lava tubes, because lavas tend to flow away from the ridge in either SE or NW directions. Resulting lava flows tend to be elongated in a direction perpendicular to the length of the ESRP and to the general direction of groundwater flow. Observed strong anisotropy and northwest–southeast elongation of the spatial correlation structure of bulk horizontal hydraulic conductivity may be related to the greater interconnectivity of rubble zones, lava tubes, and fracture systems in the direction parallel to the long dimension of lava flows (Welhan and Reed, 1997).

The rubble zones between lava flows provide the main porosity and permeability in the aquifer and the vadose zone, but platy fracture zones, columnar joints, vesicles, and vent facies pyroclastics and shelly pahoehoe also contribute. Conceptually, lava tubes could also be important, but little information exists to quantify their contribution. Groundwater takes advantage of all the open spaces, seeking out interconnected pathways for flow. Vadose zone water often takes tortuous paths to

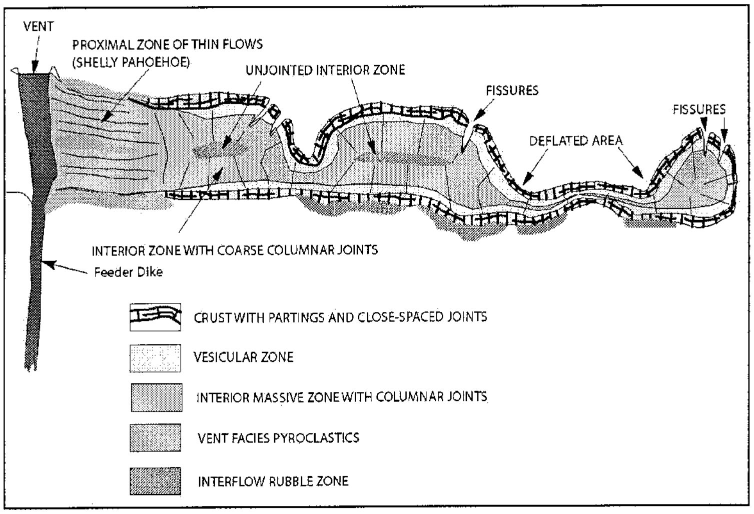

Fig. 3. Typical eastern Snake River Plain basalt lava flow showing features that impart permeability.

reach the aquifer, perching on massive flow interiors and sediment interbeds. In one area, documented lateral flow of >1 km occurs as it seeks a way downward to the water table (Nimmo et al., 2002). In other areas, tracer tests (Nimmo et al., 2002) and well response during recharge events along the Big Lost River show that water percolates directly downward to the aquifer. Temperature logs of some wells show that constant atmospheric exchange extends downward to the water table (60–200 m) in some areas where fractures and rubble zones provide interconnected pathways to great depth (Smith et al., 2004).

The rubble zones and fractures of lava flows contribute to infiltration of silty and clayey surficial and inter-

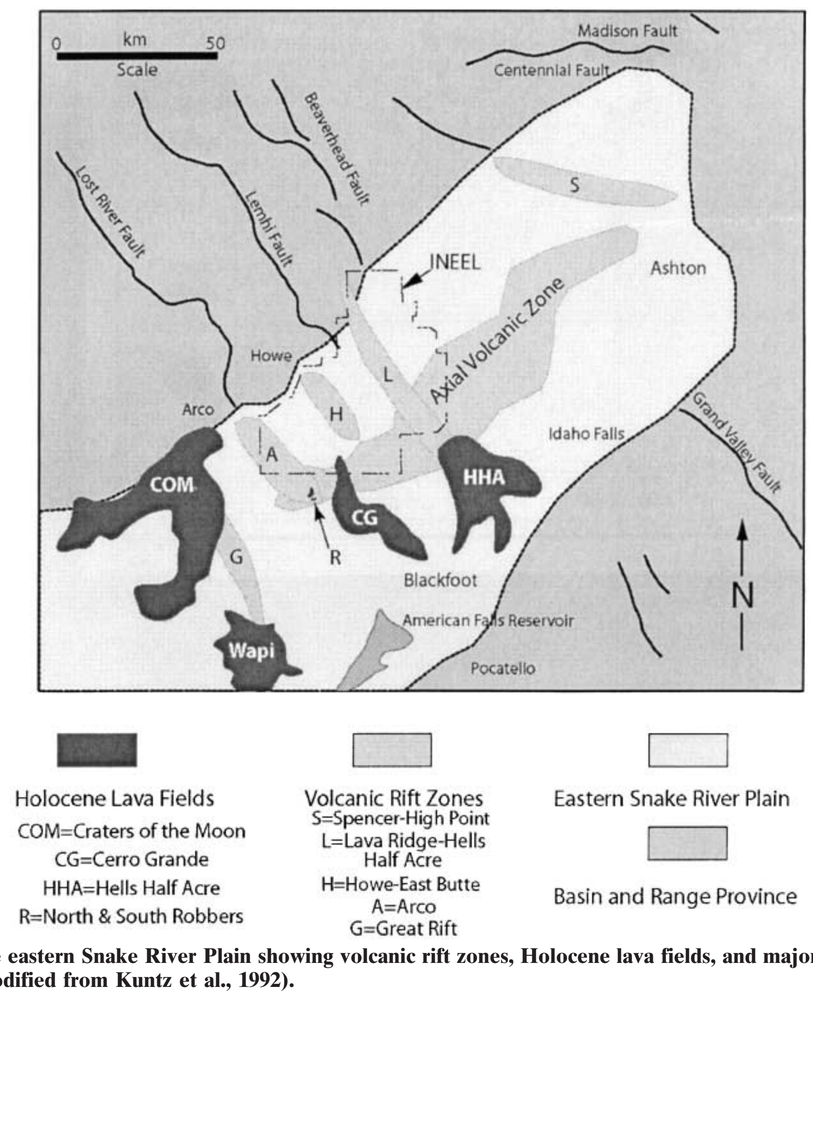

Fig. 4. Regional map of the eastern Snake River Plain showing volcanic rift zones, Holocene lava fields, and major faults in the adjacent Basin and Range province (modified from Kuntz et al., 1992).

bed sediments, which commonly fill voids and decrease porosity and permeability at depth. Such infiltrated sediment is commonly observed in drill cores just beneath interbeds, but can also occur as fillings in fractures in parts of the section remote from obvious interbed or surficial sediment source. Sediment infilling material has the effect of retarding groundwater flow and may also contribute to retardation of contaminant migration due to cation exchange and/or sorption processes.

CONSEQUENCES OF SUBSIDENCE OF THE ESRP

The ESRP surface has subsided about 1 km relative to adjacent regions after passage of the Yellowstone hotspot from beneath the area at about 4 million years ago (Smith et al., 1994). This subsidence has had strong effects on the evolution of the Snake River Plain aquifer and vadose zone. The subsiding elongate basin collects detrital sediments from surrounding mountains and makes space for the accumulation of the basalt–sediment sequence to the present thickness of 1 to 2 km (Hackett and Smith, 1992). The Snake River Plain aquifer is confined to the upper part of the basalt–sediment sequence, and the boundary of the ESRP itself defines the boundary of the aquifer. The vadose zone ranges in thickness from 0 to about 300 m, and the thickness of actively flowing water in the aquifer ranges up to about 400 m (Morse and McCurry, 2002; Smith et al., 2004).

Subsidence of the Plain is not uniform in time and space, and differential subsidence influences the accumulation of sediments and the distribution of sediment facies and interbeds. The Big Lost Trough (Geslin et al., 2002; Bestland et al., 2002; Mark and Thackray, 2002) in north-central and northeastern parts of the INEEL has been a persistent low area, and consequently, a greater proportion of sediments have accumulated and been preserved there. During the Pliocene and early Pleistocene, the climate was wetter and the water table was higher than it is today, leading to development of pluvial lakes and accumulation of laminated lacustrine silts and clays, which are preserved as interbeds in the basalt lavas (Bestland et al., 2002). Drying conditions in late Pleistocene led to deposition of playa deposits, eolian sands and silts, and weak paleosols. Coarser-grained alluvial channel deposits form sinuous stringers through the silts and clays and may provide pathways for preferential groundwater flow (Mark and Thackray, 2002). Sediment interbeds in this area of the ESRP comprise 50% or more of the stratigraphic column, and interbed thicknesses of up to 100 m occur in some places.

Along the major streams, such as the Big Lost River within the INEEL and the Snake River along the southern boundary of the ESRP, alluvial surficial deposits and interbeds of gravels and silty sands are common. The deposits are commonly clast-supported gravels with most interstitial spaces filled with silt and sand. Sandy and silty layers occur within the gravels. Gravel thicknesses of 10 to 20 m are common, and the linear, sinuous forms of these deposits may serve as preferential groundwater flow paths. The pathways produced by these coarse channel deposits between basalt layers are analogous to those described by Mark and Thackray (2002) within the fine-grained silts and clays of the Big Lost Trough.

Elsewhere in the ESRP, sediment accumulation is minor, sediment interbeds are thin (a few meters or less), and eolian and alluvial sediment types predominate. On the volcanic highlands (Axial Volcanic Zone and the volcanic rift zones), eolian silts (loess) are the major surficial sediment and interbed materials. Some local closed basins in the highland areas collect enough runoff during precipitation and snowmelt events to acquire playa characteristics, with deposition of silty sediments and pedogenic precipitation of caliche.

Subsidence of the ESRP has had a long-term effect on the amount of recharge to the aquifer. Before subsidence commenced, the aquifer had a relatively small drainage area. As the surface of the ESRP subsided, the boundaries of the drainage basin for the ancestral aquifer expanded outwards. The heads of streams that originally flowed away from the ESRP were captured and began to flow inwards (toward the ESRP) to provide surface and underflow recharge (Ruppel, 1967; Rodgers et al., 1991; Bobo, 1991; Fritz and Sears, 1993). As subsidence progressed, the drainage basin for the aquifer and for streams flowing into the ESRP grew continually larger. The present drainage divides (Fig. 2) of the northwest-trending Big Lost River, Little Lost River, and Birch Creek have migrated progressively northward from positions near the margin of the ESRP to their current positions 70 to 90 km north of the margin.

The persistent growth of the Axial Volcanic Zone by prolific volcanism prevents surface water in the Big Lost River, Little Lost River, and Birch Creek from flowing to the Snake River along the south margin of the Plain (Fig. 2). Instead, these streams flow into closed basins along the northern margin of the Plain in the INEEL area. Intermittent stream flow from these three drainages either infiltrates to the aquifer or escapes through evapotranspiration. All three drainages contribute substantial underflow recharge to the aquifer whether or not there is sufficient water for surface flow.

Because the ESRP is subsiding, streams are aggrading and sediments and basalts are continually accumulating. Basalts and sediment interbeds within the vadose zone and aquifer were formed at the surface and subsequently buried by younger basalts and sediments. In contrast to down cutting and emergent conditions typical of most upland areas, rocks in the vadose zone are not overcompacted and most of them do not have a history of being saturated or altered and mineralized by groundwater. Much of the interstitial calcite observed in vadose zone and aquifer basalts and sediments is caliche deposited by soil-forming processes before burial by younger rocks. The only postemplacement processes to which they have been subjected are slight to moderate weathering and pedogenesis in most areas of the ESRP. During Pleistocene glacial cycles, the water table was higher and caused temporary saturation of parts of the present vadose zone, but no alteration or mineralization attributable those higher water levels has been documented.

CONSEQUENCES OF CRUSTAL EXTENSION OF THE ESRP

Because the ESRP lies within the northern Basin and Range province (Fig. 1) it experiences the same northeast–southwest-directed minimum horizontal compressive stresses. The resulting anisotropy in horizontal stresses affects the aquifer and vadose zone. In situ crustal stress plays an important role in development of anisotropy in hydraulic conductivity of fractured rock aquifers such as the Snake River Plain aquifer (Finkbeiner et al., 1997; Ferrill et al., 1999), even if the orientations of fractures is random. Northwest-striking vertical fractures are oriented optimally for dilation in the ambient stress field, and those that strike northwest and dip moderately in either direction are optimally oriented for shear displacement (Moos and Barton, 1990; Jackson et al., 1993). Fractures oriented optimally for either dilation or shear have enhanced fracture permeability. Fractures of other orientations are under normal compression, causing them to tend to close. The overall effect is to hamper the southwestward flow of groundwater and to abet northwestward or southeastward flow in fractures, even in a situation in which the regional gradient of the potentiometric surface is toward the southwest and the overall distribution of fracture attitudes is random. This effect could help to explain the observed southward and southeastward movement of contaminants in some areas of the INEEL (Arnett, 2002).

Another effect is less direct, but still important in some places. The in situ stress condition dictates the orientation of basalt dikes that transmit magma from the mantle to the surface of the ESRP, and of extensional structures in the volcanic rift zones that overly them. Because of the northeast to southwest-directed extension, the orientation of dikes is northwest striking and vertical, and volcanic rift zones are northwest trending, perpendicular to the axis of the ESRP. Although exposures of dikes on the ESRP are extremely rare (due to the subsidence and continual resurfacing of the Plain), analogy with dikes in similar environments in Iceland and Hawaii (Smith et al., 1996) suggest that their thickness is 1 to 2 m and that they are less transmissive than the basalt lava flow sequence into which they are intruded. In addition, regions directly above the upper tips of propagating dikes are put into tension, and northwest-trending open fissures form (Smith et al., 1996). Unlike the columnar jointing fractures that are restricted to individual lava flows, these open fissures and the accompanying dikes are through-going and crosscut numerous lava flows in the aquifer and vadose zone. Although mapping at the surface (Kuntz, 1992; Kuntz et al., 1994) shows that these structures are concentrated in volcanic rift zones and in the Axial Volcanic Zone (Fig. 4), they may be more widespread in the subsurface due to migration of volcanic rift zones and volcanic centers with time (Kuntz et al., 2002). Both the dikes (which act as vertical impediments to groundwater flow) and the open fissures (which act as ready northwest- and southwest-striking transmissive zones) tend to induce water flow in a direction perpendicular to the general gradient of the water table. The fissures also affect the vadose zone by providing pathways for deep barometric pumping of air, potentially drying the rocks in places and causing lateral migration of fluids. The fissures are also conduits for deep penetration of fine-grained surficial sediments (silts and clays), which can then move laterally into rubble zones between lava flows and into openings within the lava flows. Such sediment infiltration may affect the transmission of water and the retardation of contaminants in the aquifer and vadose zone.

CONSEQUENCES OF THE HIGH HEAT FLOW OF THE ESRP

The conductive crustal heat flow beneath the Snake River Plain aquifer is among the highest on the continent (110 mW m^{-2}), but the heat is intercepted by cold, rapidly flowing groundwater, and the heat flow at the surface of the Plain is very low ($<$20 mW m^{-2}) (Brott et al., 1981; Blackwell, 1989). The high heat flow beneath the aquifer can be detected and measured only in a few deep wells that penetrate beneath the base of the aquifer (Fig. 5). The temperature profiles in these wells provide information about aquifer geometry. Temperatures from the water table are nearly isothermal to the base of the aquifer (200–600 m), where an inflection to a steep conductive gradient occurs. The temperature inflection represents the transition from permeable rocks with rapidly flowing groundwater above, to impermeable rocks in which very little water movement occurs. Drill core from some of the deep wells shows that the temperature inflection also marks the upper limit of hydrothermal alteration and mineralization of the basalts (Morse and McCurry, 2002). The position of the base of the aquifer is a zone of dynamic equilibrium between cold aquifer waters above and geothermal fluids and conductive heat from below. If permeability is lowered in part of the aquifer, due to high content of fine-grained sediment interbeds for example, the geothermal system encroaches upward and begins sealing porosity with precipitated minerals and altering primary minerals in the basalt and sediment.

One of the major effects of the high heat flow on the aquifer is the process of sealing permeability and causing the aquifer to be confined to those unaltered and permeable rocks in the upper part of the basalt sequence. The upper boundary of the aquifer (the water table) is controlled by the amount of infiltrating water from the watershed and varies with climatic variability and drought cycles. Major variation (several tens of meters) in elevation of the water table and thickness of the aquifer are likely to have been associated with waxing and waning of glacial cycles in Pleistocene time. More moderate variations (several meters at most) accompany modern seasonal- and decadal-scale cycles of precipitation variability. In the present climatic conditions the amount of recharge water is sufficient to maintain water table

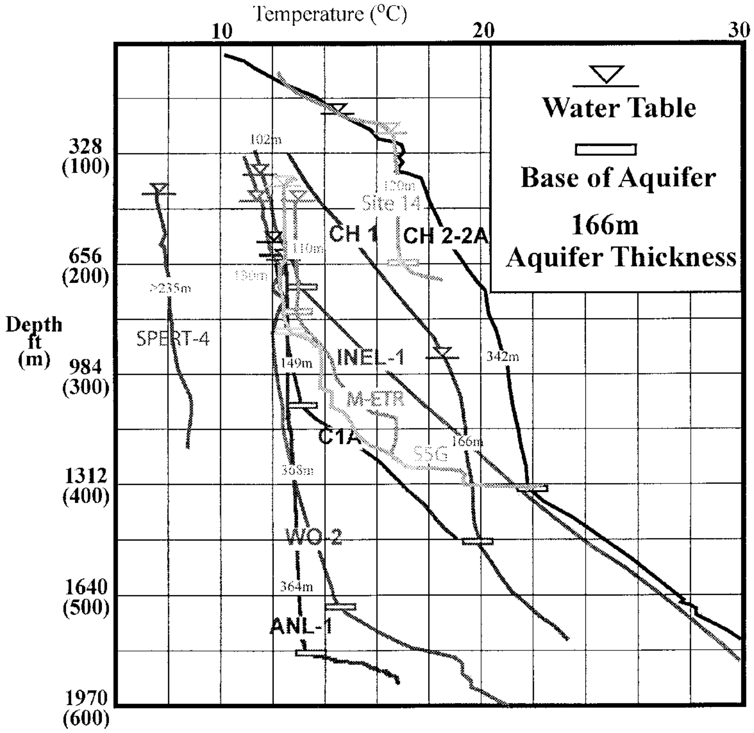

Fig. 5. Temperature profiles of deep wells at and near the INEEL.

elevations of 1300 to 1400 m above mean sea level, and this leads to a thick vadose zone in the INEEL area ranging from about 60 m in the north to more than 300 m in the south. Wetter conditions associated with Pleistocene glacial cycles may have saturated the lower parts of the present vadose zone, and if so, any effects have not been documented.

Another important effect of the high heat flow is the warming of the aquifer water as it moves from recharge areas on the Yellowstone Plateau and other highlands around the Plain to discharge areas along the Snake River at Thousand Springs (Fig. 1 and 4). Recharge water enters the system from snow melt at high elevation and usually has a temperature of 5 to 7°C. It is slowly warmed during transit to discharge areas and leaves the aquifer at around 16°C. The general heating was modeled by Brott et al. (1981) and is illustrated in Fig. 6. In addition to the gradual warming with distance from recharge areas, there are local areas of anomalously warm water and others with anomalously cold water. The warm zones represent either areas of low permeability in which water moves slowly and is heated by heat transfer from below or areas of vigorous geothermal input that overwhelms the cold aquifer waters above.

The aquifer water in the INEEL area has been heated during its transit from up-gradient areas to around 11 to 12°C. The temperature of recharge water from drainages north of the INEEL is 7 to 9°C, and the temperature difference provides a way to trace groundwater pathways as local recharge water moves into the aquifer. Several preferential pathways of groundwater movement have been identified by temperature distribution in the aquifer on and near INEEL (Smith et al., 2004). Some of these pathways are revealed by narrow plumes of cold recharge water penetrating warm zones in the aquifer.

The local zones of anomalously warm (up to 18°C) and cold (7–9°C) water in the aquifer affect the temperature and temperature gradient in the vadose zone above. In anomalously warm areas, the average temperature gradient from the surface to the water table is higher than normal, and in the cold zones the gradient can be very low to negative. Such variations in temperature gradient may have implications for fluid and contaminant transport in the vadose zone.

CONSEQUENCES OF THE CONTEMPORARY TOPOGRAPHY AND LAND USE

The present high elevations of the Yellowstone Plateau and the Basin-and-Range mountains surrounding the ESRP provide a strong orographic uplift for winter storms from the Pacific. Consequently, these areas accumulate large winter snowpacks that furnish abundant water to maintain a fast-moving and highly productive aquifer. The drainage basin also has rugged topography and is remote from large cities and major industrial facilities; much of it under the control of the National Park Service and the U.S. Forest Service. The water entering the aquifer from these areas is mostly pure and uncontaminated, contributing to the high quality of the water. This is significant because it allows ready identification of most sources of groundwater contamination and makes the definition of contaminate plumes realtive straightforward. The major threats to water quality in the aquifer are agricultural activities (fertilizer and pesticide use, reinjection of runoff from irrigation activities,

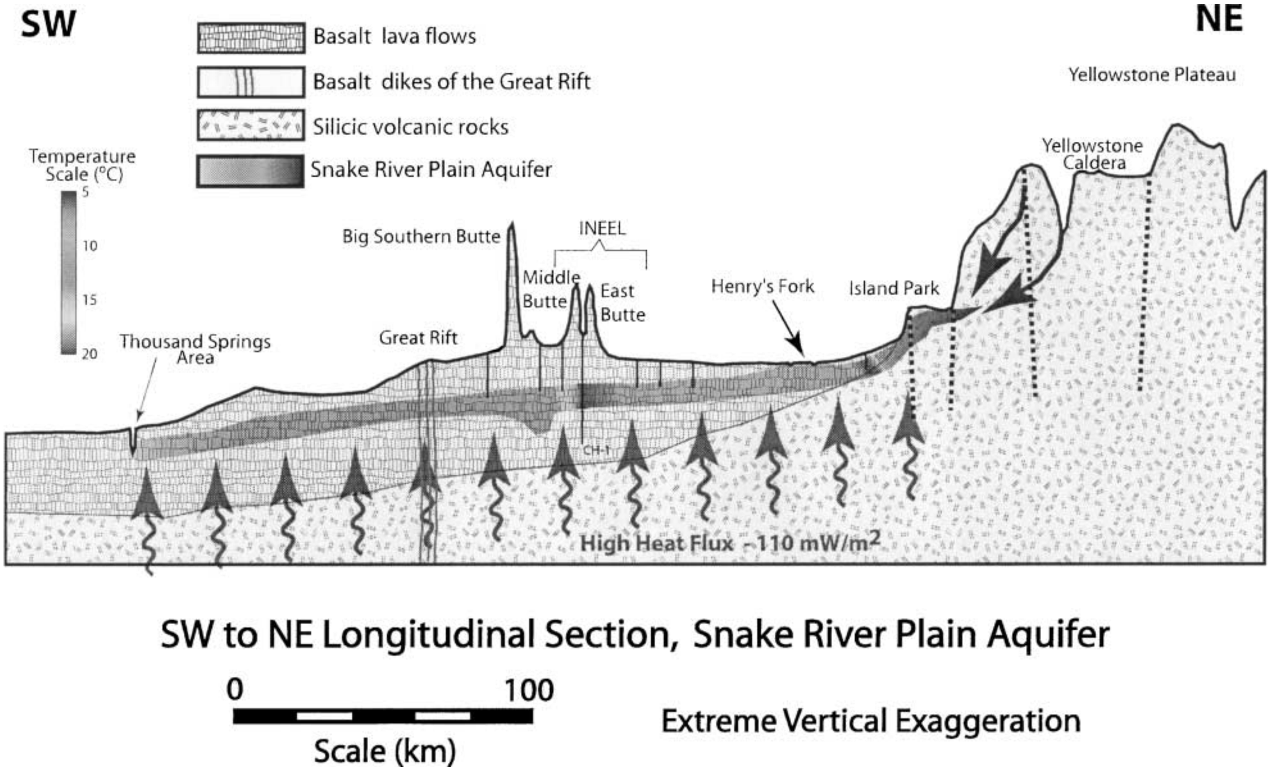

Fig. 6. Longitudinal section of the aquifer showing gradual warming caused by high conductive heat flow from beneath the eastern Snake River Plain.

and disposal of feedlot and dairy wastes), some of the larger municipalities (urban runoff, pipeline and tank leakage, spills), and some past waste disposal practices at the INEEL (injection wells, burial of wastes, storage of reprocessing wastes, and infiltration ponds).

CONCLUSIONS

The properties of both the Snake River Plain aquifer and its overlying vadose zone are strongly influenced by tectonic and volcanic processes that formed the ESRP. The size, shape, water volume, flow rates, and temperature distribution of the aquifer are determined by disparate processes of lava flow emplacement, surface subsidence, sediment accumulation, extensional stresses, heat flow from the deep crust, and meteorological conditions associated with high elevations surrounding he ESRP. The complex nature of the vadose zone results from many of these same processes, and the resulting abrupt spatial variability in physical and chemical properties of rocks and sediments in the subsurface.

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