Chapter 2

GROUNDWATER FLOW ON THE APPALACHIAN PLATEAU OF PENNSYLVANIA

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Introduction

Acid mine drainage (AMD) results from the oxidation of sulfide minerals in the strata disturbed by mining. The products of this reaction. acidity, sulfate, and iron, (see Chapter 1) are dissolved in and transported by groundwater. Consequently, in addition to understanding the geochemical nature of the enveloping strata, prediction and prevention of AMD requires an understanding of how groundwater flows through the mine and adjacent area. Groundwater flow in the coal fields of the Appalachian Plateau of Pennsylvania can be viewed from the scale of the entire groundwater flow system down to the detailed level of flow through aquifers and aquitards on a mine site. This flow system can be influenced by many factors including stratigraphy and lithology, structural geology, tectonic stresses and their cumulative effects on rock-mass integrity, geomorphology, topography, and any substantial man-made influences such as mine subsidence related fracturing.

An understanding of the groundwater flow regime is essential to the design of site-specific monitoring plans to determine the impact of mining on groundwater quality and the hydrologic balance. The more thorough the understanding of this system, the more efficient this plan can be, maximizing the information which can be obtained and minimizing the costs of monitoring. This knowledge also gives the groundwater chemistry of the site meaning by providing a frame of reference regarding a given monitoring point's location and relative importance within the groundwater flow system. Absent this understanding, considerable time and money can be wasted by the poor placement of groundwater monitoring wells and the subsequent collection of meaningless water quality data. Although groundwater chemistry can be used to help characterize the flow system, it must be integrated within a broader effort intended to define groundwater potential and hydraulic gradient. The importance of understanding groundwater flow at a site is emphasized by Earl (1986) who suggested that for groundwater investigations, 90% of the budget should be used to define the flow system, and the remainder for water chemistry analysis.

Many of the principal concerns in the review of mine permit applications - prevention of adverse impacts to water resources, prevention of postmining AMD, and the prediction of impacts to water supplies, require knowledge of premining groundwater conditions and an understanding of how mining will change these conditions. An environmentally prudent mine-site design relies on accurate characterization of the premining groundwater flow regime. See Chapters 15 and 16 for discussions on the management of mine-site water.

Climate

AMD problems in the northern and central Appalachian coal mining region far outstrip problems in other coal producing regions of the United States (see Chapter 8). Some of the regional differences are related to climate. The process of rock weathering is strongly influenced by temperature and by amount and distribution of precipitation (Hem, 1992). AMD formation is most severe in humid areas with moderate rainfall where rapid oxidation and dissolution of exposed minerals can occur. Therefore, any discussion of AMD prediction and prevention in Pennsylvania must be considered within the context of Pennsylvania's climate. Observations and conclusions made in Pennsylvania may not be relevant to other areas of the country due to climatic differences. For example, the interplay between pyritic oxidation and the dissolution of carbonate minerals may be very different in semi-arid or arid climates.

Pennsylvania has a humid climate with precipitation distributed relatively evenly throughout the year. The amount of rainfall and snowfall vary over Pennsylvania. The most recent precipitation normals from the

National Climatic Data Center for Pennsylvania for the time period of 1961 to 1990 show the lowest annual average amount to be 32 inches near Covington, Tioga County, and the highest amount of precipitation to be 49.5 inches is near Tamaqua, Schuylkill County. Figure 2.1 shows precipitation isolines which range from 36 inches to 50 inches in Pennsylvania. Snowfall isolines (Figure 2.2) illustrate average snowfalls of 20 to 100 inches throughout the state.

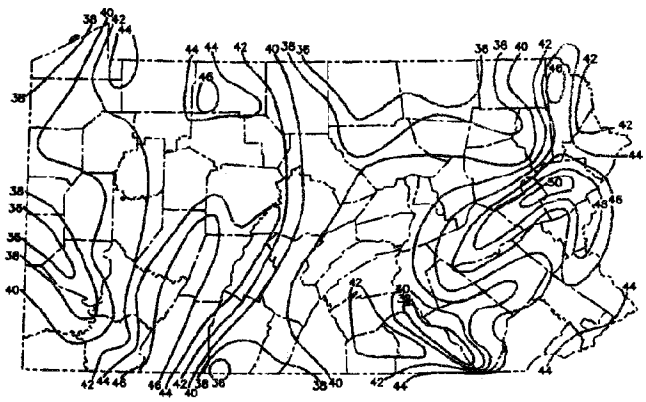


Figure 2.1 Mean annual precipitation, inches. (Based on the period 1931-1960)

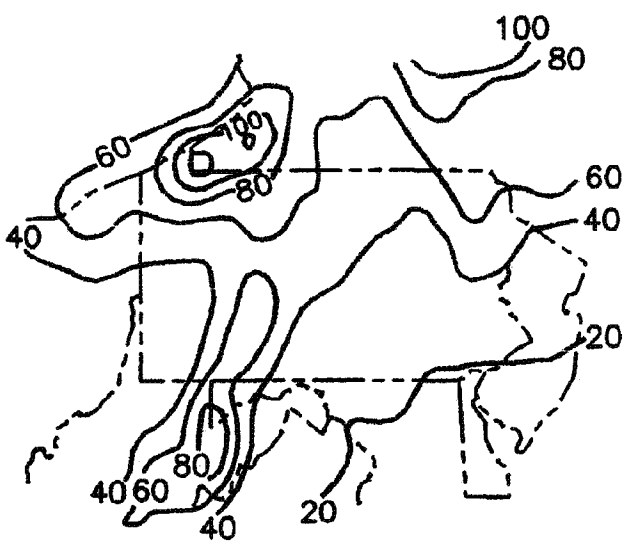


Figure 2.2 Total snowfall (inches), 50th percentile, seasonal.

Pennsylvania's temperatures vary greatly across the state during the year. The average temperature for the most recent time period (1961-1990) is about 48° F. The temperature fluctuations are revealed by examining the temperatures during the normally coldest month (January) and the normally hottest month (July). Figures 2.3 and 2.4 show the mean minimum and the mean maximum temperatures for January, while Figures 2.5 and 2.6 show the mean minimum and the mean maximum temperatures for July. This information is from data published by the National Climatic Data Center and covers the time period of 1931-1960. During a more recent time period (1961-1990) a similar pattern appears to hold.

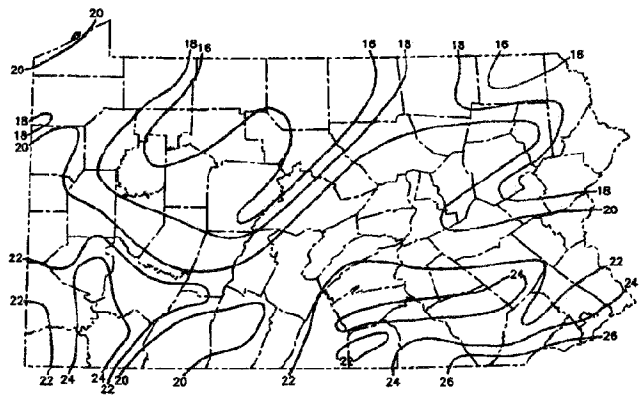


Figure 2.3 Mean minimum temperature (°F), January. (Based on period 1931-1960)

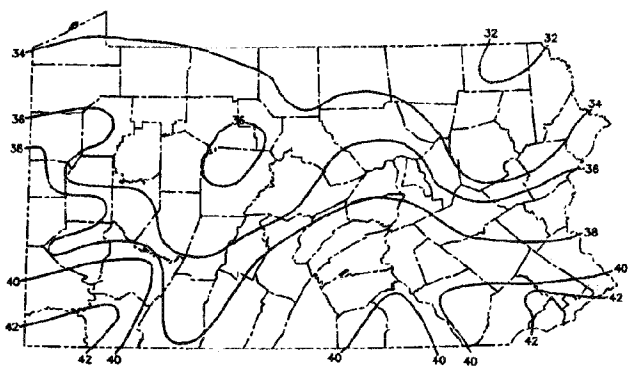


Figure 2.4 Mean maximum temperature (°F), January (Based on period 1931-1960)

In Pennsylvania approximately 12 to 15 inches annually, or about one third of the average precipitation, infiltrates to the groundwater system. Evaporation and transpiration account for about 20 inches returning to the atmosphere annually (Becher, 1978). The remaining precipitation directly runs off to surface waterways. Of course, these are averages. The actual amounts vary from place-to-place depending on geology, soils, vegetation, topography, and from year-to-year. Areas with readily permeable soils, porous bedrock, or closed surface depressions generally have higher infiltration rates (Smith, 1986).

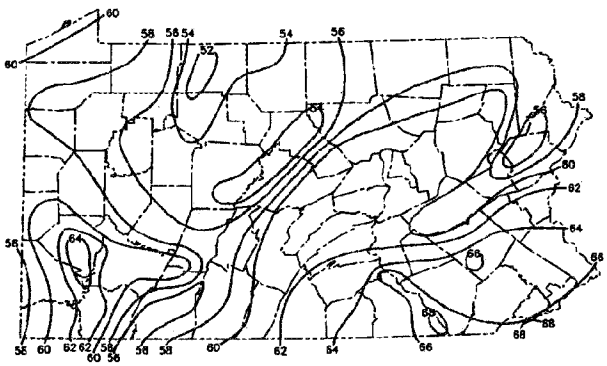


Figure 2.5 Mean minimum temperature (°F), July. (Based on period 1931-1960)

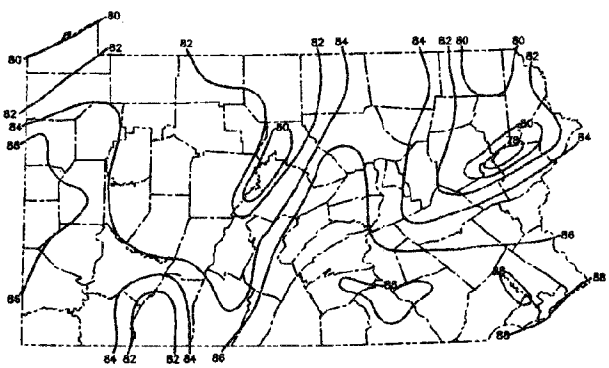


Figure 2.6 Mean maximum temperature (°F), July. (Based on period 1931-1960)

There are several aspects of this "hydrologic budget" which are important to consider. First, because precipitation exceeds evapotranspiration in all of Pennsylvania, virtually all areas, unless capped with some impervious material, are subject to infiltration and produce groundwater. In this respect, there is no such thing as a completely dry mine site - one that does not produce groundwater drainage. If 12 to 15 inches of annual groundwater discharge were expressed as an average flow rate, every acre of land would produce a flow of 0.62 to 0.77 gallons per minute (alternately, one hectare would produce 5.8 to 7.2 L/sec). Also, groundwater recharge does not occur uniformly throughout the year. Recharge is greatest during the non-growing season when there is minimal uptake of water by plants and evapotranspiration is minimal (Smith, 1986).

Groundwater Flow

Groundwater flow is largely controlled by three factors - the distribution and quantity of recharge to the flow system, surface topography, and the hydraulic conductivity of the material through which the groundwater flows. These factors may in turn be affected by a host of other elements - soils, climate, lithology, and geologic structure. But, at a fundamental level, the groundwater flow system can be reduced to these principal elements. Mining, both surface and underground, can drastically alter each of these factors. Groundwater recharge rates can be altered depending on how the surface material is handled and how the site is revegetated. Surface topography can be altered depending on the final reclamation contours (see Chapter 12). Hydraulic conductivity of the overburden, which must be removed and replaced at surface mines and which can be collapsed and fractured above underground mines, is dramatically and permanently altered by the mining process. Additionally, mine spoil may be more conductive than parent material by several orders of magnitude. (The hydraulic properties of mine spoil are discussed in detail in Chapter 3.)

Hydraulic Head

Groundwater flow is driven by differences in hydraulic head. Groundwater flows from areas with higher hydraulic head to areas with lower hydraulic head. In an unconfined aquifer, the elevation of the water table surface can be used to determine distribution of hydraulic head and to infer the direction of groundwater flow. The hydraulic head at any point can be determined by the water level in a piezometer. Determination of the hydraulic head at a sufficient number of locations can be used to create a potentiometric map showing lines of equal potential (head). This map can be used to determine the direction of groundwater flow - from high potential to low potential.

Determination of the hydraulic head in an unconfined system or in multiple aquifers at the same location indicates whether the vertical component of groundwater flow is down or up (recharge or discharge area). Determination of the hydraulic head at a sufficient number of locations areally and vertically can be used to obtain a three-dimensional determination of groundwater flow. The flow directions within a local flow system are frequently estimated by assuming they approximate the topography, coupled with spring (discharge area) and, possibly, well (water level) mapping. Static water levels in wells can be used as estimates of hydraulic head in single aquifer wells. These approximations are reasonably accurate for fairly homogeneous bedrock flow systems but may be in error for areas with large permeability contrasts.

The hydraulic gradient and anisotropy of the transmitting medium control the specific flow paths between recharge and discharge areas. In homogeneous, isotropic media groundwater flows perpendicular to equipotential lines. Anisotropy can result in flow which is oblique with respect to equipotential lines (Fetter, 1981).

Static Water Level

When a borehole is drilled, the water level will stabilize at the static water level which represents the composite hydraulic head of the open interval of the well. Figure 2.7 illustrates the relationship between a producing aquifer (aquifer 1), a thieving aquifer (aquifer 2), their piezometric surfaces, and the resulting static water level in a recharge area well in Mercer County. Each of the aquifers has a different hydraulic head, represented by the potentiometric surface. The static water level is the composite of the hydraulic heads of the two aquifers.

In recharge areas groundwater movement has a downward component. Therefore, as a borehole is deepened and additional aquifers are penetrated, the composite head decreases resulting in a drop in the static water level. In discharge areas the reverse is true, as deeper aquifers are penetrated the composite head increases and the static water level rises (Bennett and Patten, 1960; Carswell and Bennett, 1963; Poth, 1963).

An erroneous interpretation of the significance of the static water level is that it indicates the water table. The water table is the boundary between saturated and unsaturated media under atmospheric pressure. The depth of the water table cannot be determined by the static water level unless the borehole is shallow and is just deep enough to encounter standing water at the bottom, generally no more than 6 feet (1.83 m) into the zone of saturation (Saines, 1981). Because the static water level indicates the composite head of the open interval in the borehole, immediately upon further deepening of a borehole, the static water level no longer reflects the water table. Interpreting the static water level to be the water table infers that the water table at the locale of a flowing well is above the land surface. The often used terms "local" and "regional" water table should not be confused with local and regional flow systems or aquifers. The "local" water table, or perched water table, is one with unsaturated media beneath it. It is usually associated with a perched aquifer.

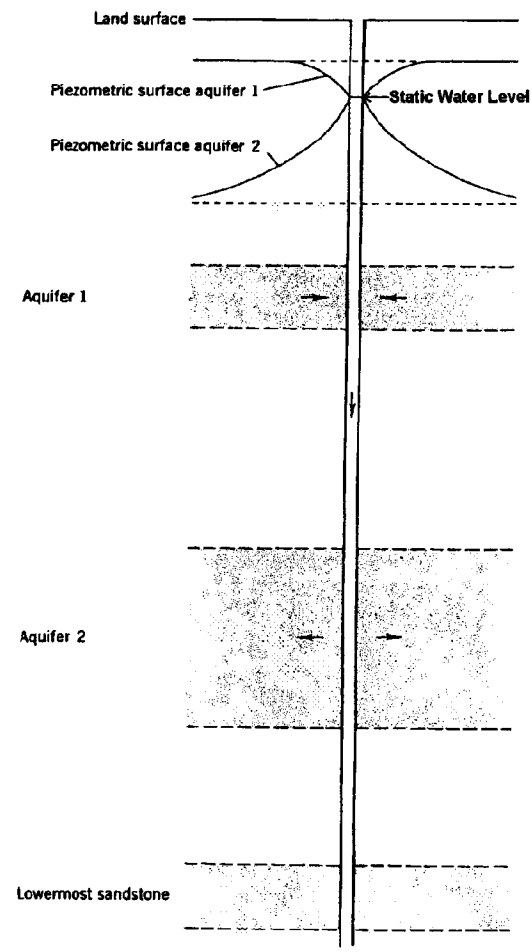


Figure 2.7 Diagram of well through two aquifers in Mercer County, PA (modified from Bennett and Patten, 1960).

Another common misconception is that the static water level indicates the level at which water is entering the borehole. That infers that the pressure head in the aquifer is atmospheric, which is true only at the water table. It also infers that the level that water is entering a flowing well is above the land surface. In Figure 2.7, neither the static water level of the well, nor the potentiometric surface, is at the level of either aquifer.

Hydraulic Conductivity

Porous media may be characterized by their hydraulic conductivity (permeability), described in general terms as either primary or secondary. Primary permeability refers to the intergranular spaces of the transmitting medium. It may be dominant in unconsolidated sediments, but is less important in consolidated bedrock of the Appalachian Plateau of Pennsylvania (Figure 2.8). A significant amount of

permeability (and porosity) in this system is secondary (Schubert, 1980). Bedding-plane partings, stress-relief fractures, zones of fracture concentration, tectonic joints, and faults serve as secondary pathways for fluid migration. Fracture zones (high concentrations of near vertical joints) are especially permeable (Lattman and Parizek, 1964). Where a coal serves as an aquifer, permeability is usually secondary along the cleat (Wunsch, 1993). High yielding aquifers in Pennsylvania's Appalachian Plateau tend to be mainly sandstones (Carswell and Bennett, 1963; Peffer, 1991), suggesting that primary permeability is a factor. However, it may be that secondary permeability tends to be better developed in the more brittle sandstones, than in shales (Carswell and Bennett, 1963; Peffer, 1991). Brown and Parizek (1971) and Wyrick and Borchers (1981) found secondary fracture permeability to be considerably greater than primary permeability. As a general rule, secondary permeability decreases with depth because fractures close due to overburden pressures and routinely decrease in frequency and distribution. Because the secondary permeability features are less pronounced in deeper regional flow systems, their overall transmissivity is significantly lower than in shallow aquifers in the local flow systems.

A general decrease in hydraulic conductivity and storativity with depth has been documented many times. Bruhn (1985) found that the average hydraulic conductivity within 150 feet (46 m) of the surface at a West Virginia mine site was 10^{-5} m/sec, decreasing to 10^{-9} m/sec for depths greater than 300 feet (90 m). Specific storativities for sandstone aquifers also decreased from 10^{-5} to 10^{-12}/ft. Stoner (1983), working in Greene County, PA, found the hydraulic conductivity reduced by one order of magnitude per 100 feet (30.48 m) of depth to a depth of 500 feet (152.4 m).

Brown and Parizek (1971) documented considerable differences between values of permeability determined from core analysis and those determined from pumping test data in Clearfield County, Pennsylvania. Values of permeability obtained from cores were markedly lower than those obtained from the pumping tests. This was attributed to the fact that the core analyses measured only the primary or intergranular permeability, whereas the field pumping test data combined the effects of both intergranular and fracture permeability.

Abandoned underground coal mine voids and caves, mainly in the Vanport Limestone, also serve as secondary permeability pathways. The flow is different from intergranular and fracture flow, especially if the conduits are not full. Shuster and White (1971) have described this flow in caves of carbonate terrain as

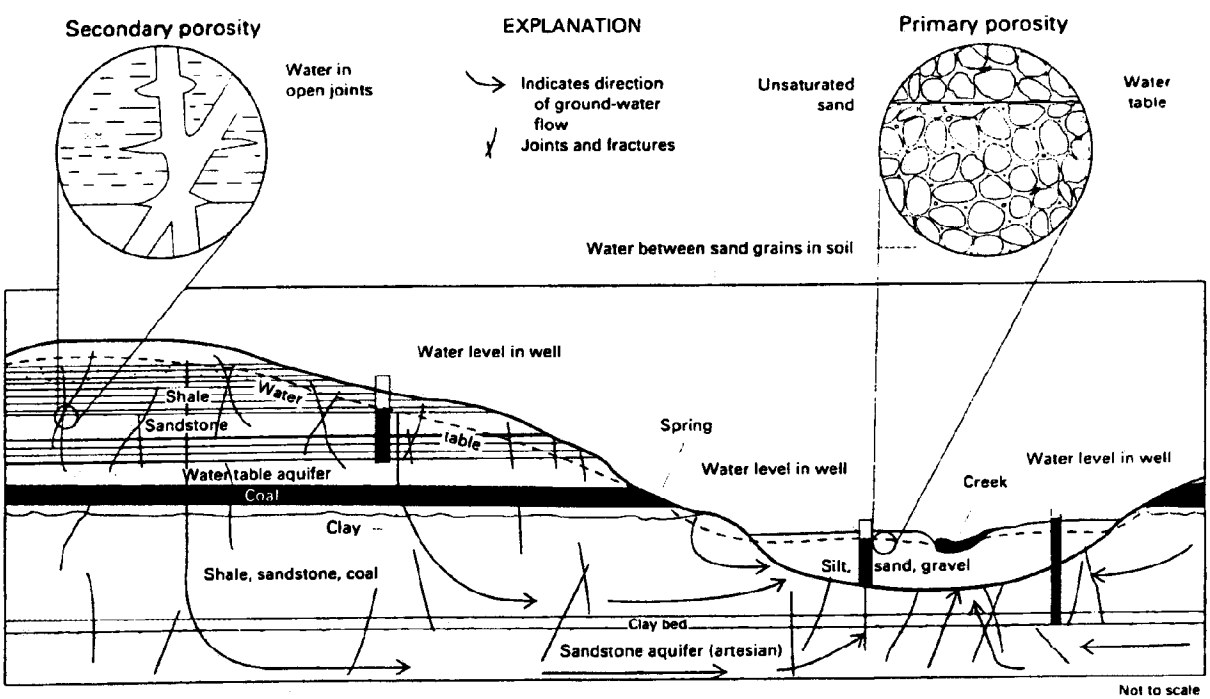


Figure 2.8 Water table aquifer (upper), confined aquifer (lower), and groundwater flow lines (modified from Hobba, 1987).

conduit flow. This term is also appropriate on the Appalachian Plateau. It is an extreme form of anisotropy in the transmitting medium. Conduit flow effectively reduces recharge to deeper aquifers and flow systems. Dispersion of groundwater via intergranular, and to a lesser extent, fracture flow, is virtually eliminated in conduit flow. As a result, contaminants are more likely to move as a slug for long distances through discrete conduits rather than being dispersed over a larger area (Freeze and Cherry, 1987).

Over large areas of the Appalachian Plateau extensive underground mining has taken place leaving, as a remnant, substantial man-made aquifers (mine-void and subsidence-fracture zones) which can have a profound influence on groundwater flow. Research by government agencies, coal producers, and major universities in the eastern coal fields has resulted in a good understanding of subsidence-fracture profiles which develop above underground mine excavations. The result is that mining associated zones of collapse, fracturing, fracture dilation, and surface disturbance can be anticipated. Various conceptual models linking strata deformation, hydraulic property changes, and impacts to groundwater uses have been put forward (Kendorski, 1994; Booth, 1986; Peng and Chiang, 1984; Hill et al.,1984; Owili-Eger, 1983; Tieman and Rauch, 1986, to name a few). Impacts to local groundwater resources vary depending on factors such as proximity to mining, mining sequence, geology, coal extraction percentage, thickness and strength of overburden, height of target coal, topography, and presence of preexisting fracture sets which may extend mining induced hydrologic impacts.

Underground mines, because of the scale of the operations, may impact the hydrology of a given locale much more profoundly than a surface mine. Underground mines have the potential to impact surface and groundwater systems on a relatively large scale. Interbasin transfer of groundwater is a common occurrence associated with underground mines. During mining the underground excavation acts as a large sink which draws in groundwater. Upon closure and flooding, it becomes a highly permeable aquifer which can permanently alter the premining flow regime both physically and chemically. Due to the open nature of the mine-void aquifer, there is a postmining transfer of the resulting mine-pool potential throughout the interconnected mine workings. This is an important factor regarding the potential for mine pool breakout since areas within and adjacent to the downdip portions of the mine workings can often realize abnormally high postmining heads comparative to premining values.

Effect of Dip on Groundwater Flow

A common misconception is that groundwater always flows downdip. Groundwater flows from recharge to discharge areas. If the discharge area is in the updip direction (from B to C on Figure 2.9), groundwater will move counter to the control of dip. Flow is governed by the relative distribution of head and not necessarily attitude of the beds.

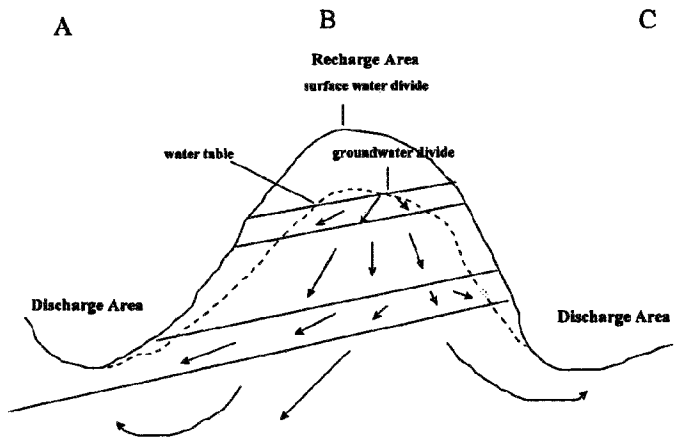


Figure 2.9 Hypothetical groundwater flow in area of dipping bedrock.

The effect that dip has on groundwater flow is that it can shift the groundwater divide updip from the surface water divide (Figure 2.9) (Earl, 1986; Shuster, 1979). Excavations, such as surface and underground mines, can serve to magnify the shifting of the groundwater divide due to their abilities to disturb hydraulic gradients and convey groundwater.

Also, bedding can have an important effect on postmining groundwater flow at surface mines. Due to the large permeability contrast between the poorly-permeable pit floor and the readily-permeable mine spoil, groundwater will flow laterally through the spoil, parallel to the pit floor, to a point of discharge at the spoil toe. Where the pit floor is inclined away from the spoil outslope, groundwater commonly impounds in the spoil to the elevation necessary to discharge either laterally or at the up-dip spoil outslope.

Pit Floor Leakage

On many surface mines, downward leakage of groundwater thorough the pit floor exceeds the rate of recharge. Where this occurs, there may be no observable discharges from the surface mine spoil, or dis-

charges may only occur during wet seasons when recharge is greater. Instead, water migrates vertically through the pit floor material at a rate sufficient to drain the overlying spoil This water ultimately reports to lower groundwater discharge zones, commonly associated with underlying coal seams.

Surface mine spoils with no apparent groundwater discharge are often erroneously thought to represent "dry sites" which do not produce drainage. But in Pennsylvania and other humid states with precipitation exceeding evapotranspiration, short of capping the site with an impermeable barrier, it is virtually impossible to prevent infiltration into the groundwater system. Instead, an apparently dry site is indicative of a leaking pit floor which transmits groundwater to underlying horizons. Leakage through the pit floor may result from the intrinsic permeability of the pit floor material, jointing or faults in the pet floor material, mine subsident fracturing, or drill hole penetration.

Fractures

Fractures and bedding-plane partings are commonly the main groundwater flow paths on the Appalachian Plateau. Near-vertical fractures (joints and faults) may cause flow to be oblique, instead of perpendicular, to the hydraulic gradient (Fetter, 1981). However, hydraulic head must decrease along the flow path. The hydraulic gradient, not the fractures, exerts the dominant control on the direction of groundwater flow. Fractures can have a direct influence on groundwater flow rate due to the generally lower frictional resistance to groundwater flow within fractures versus intergranular pores and to the fractures' role in lessening flow system tortuosity.

The following is a list of secondary permeability features which can impart significant local and/or regional controls on the flow system of the northern Appalachian coal measures:

Joints - A joint is a rock fracture along which displacement has not occurred. The joint pattern is cumulative and represents a record of all stress events sufficient to induce fractures. On the Appalachian Plateau there are two ubiquitous, orthogonal joint sets which form the *fundamental joint system.* The joint sets are the systematic joints (planar joints in shales and sandstones, face cleats in coal) and the nonsystematic joints (curved joints in shales and sandstones, butt cleats in coal). Systematic joints are the dominate set and continue across other joints; are generally perpendicular to upper and lower rock unit boundaries; are commonly perpendicular to bedding; and are <u>not</u> confined to rock type but can pass downward or upward into adjacent units. Nonsystematic joints are nonplanar; do not generally cross other joints; and commonly terminate against bedding planes (Nickelsen and Hough, 1967).

Joints from different sets and/or different lithologies have different properties. Spacing (distance between joints) and width (distance across a joint) are generally greatest in coarser, more resistant lithologies. Systematic joints may continue into adjacent beds or lithologies, whereas nonsystematic joints generally terminate against systematic joints, bedding planes, and lithologic contacts. Joints (cleats) in coal do not usually pass into adjacent lithologies. Joints in sandstones are parallel and of similar planeness as shale joints but are more widely spaced (sandstones show spacing of many meters; shales show spacing of several centimeters to many meters; coals show spacing of fractions of a millimeter to several centimeters) (Nickelsen and Hough, 1967).

Joint characteristics affect groundwater flow. Spacing suggests the number of joints available for groundwater pathways. The width indicates the ability of the joints to transmit water. Systematic joints are often continuous and can transmit water longer distances, more rapidly, than non-continuous, nonsystematic joints. Joint permeability is particularly significant, if not dominant, in shallow rock strata of western Pennsylvania (50-500 feet) (15-150 m) and may increase permeability values by ten to a thousand times when compared to intergranular permeability values taken alone (Parizek, 1971).

Stress-relief fractures - Stress-relief fractures are a fracture network unrelated in age and orientation to tectonic stresses. They are often the most transmissive part of an aquifer (Wyrick and Borchers, 1981). They were first documented during abutment characterization prior to the construction of dams throughout Pennsylvania, Ohio, Kentucky, New York, West Virginia, and other areas worldwide where similar flat-lying sedimentary rocks were drilled or excavated during foundation work (Burwell and Moneymaker, 1950; Ferguson, 1967, 1974; Ferguson et al., 1981). Stress-relief fractures include vertical fractures parallel to valley walls, caused by the release of stress as lateral rock support is removed during erosional downcutting. Slumping along the fractures on both sides of a valley cause compression in the center of the valley, resulting in thrust faults, bedding-plane partings, arching, and

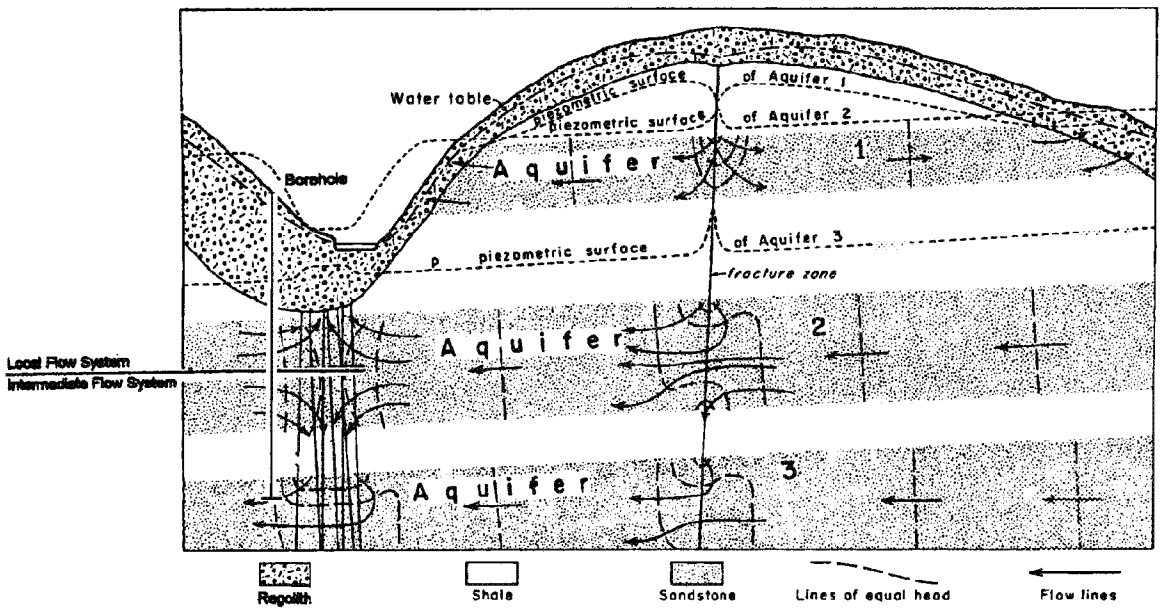


Figure 2.10 Idealized pattern of groundwater flow in and beneath a "hydrologic island" (modified from Poth, 1963).

vertical extension fractures above arches (see Figure 2.16). Studies by Wyrick and Borchers (1981), Kipp and Dinger (1987), and Davis (1987) describe a highly permeable, valley-related, shallow flow system which consists of interconnected valley-wall and valley-floor fracture sets. Sections of the valley-floor portion of the subsystem can become artesian due to the presence of alluvial clay which generally occurs mid-valley and can serve as a confining layer. Sames and Moebs (1991) describe the character and extent of valley-wall, stress-relief fractures as follows:

> "...Stress-relief joints occur in the shallow overburden where surface slopes are steep, with the greatest frequency and degree of weathering in mine roof within 200 feet laterally of coalbed outcrop. They are usually discontinuous along strike, and decrease in frequency and degree of weathering to about 700 feet from the outcrop and under 300 feet or more of overburden, as the effects of stress relief gradually disappear."

Zones of fracture concentration - Fracture zones (see Figure 2.10 and 2.11) are relatively restricted areas where numerous fractures dissect the rock mass. The degree of interconnectedness varies as depicted on Figure 2.12. Fracture zone widths of 7 to 12 meters are common in the central Appalachian region (Gold, 1980). Depths range into the hundreds of meters. The hydrologic impacts of a fracture zone can be profound. An entire discipline has evolved around locating fracture zones by mapping their surface expressions (fracture traces and lineaments) and exploiting their relatively high permeabilities for the development of water supply wells. Fracture zones are often valley related phenomena where they can become integrated with valley stress-relief fracture systems. These features can, and do, serve as major conduits for groundwater. In zones of relatively intense fracturing, hydraulic conductivities are often several orders of magnitude higher than in unfractured rocks, resulting in fracture-dominated flow (Schubert, 1980).

Bedding-plane partings - Inherent weaknesses in rock arising from thin bedding (laminations), fissility and/or lithologic contacts often are zones which will provide avenues for groundwater migration. Schubert (1980) discusses the following observation from a drilling project in West Virginia, ... "The only large quantities of water encountered during drilling originated from fractures or from contacts between formations. Shales often supplied more water than did the coarse but well-cemented sandstones, apparently because of slight fracturing which permitted good lateral movement of water".

Fault zones - A fault is a fracture or fracture set along which there has been displacement. Faults can be important conveyers of groundwater relative to the surrounding unbroken rock mass and are common in some of the more eastern portions of the Plateau, near the Allegheny Front, where stresses were maximized

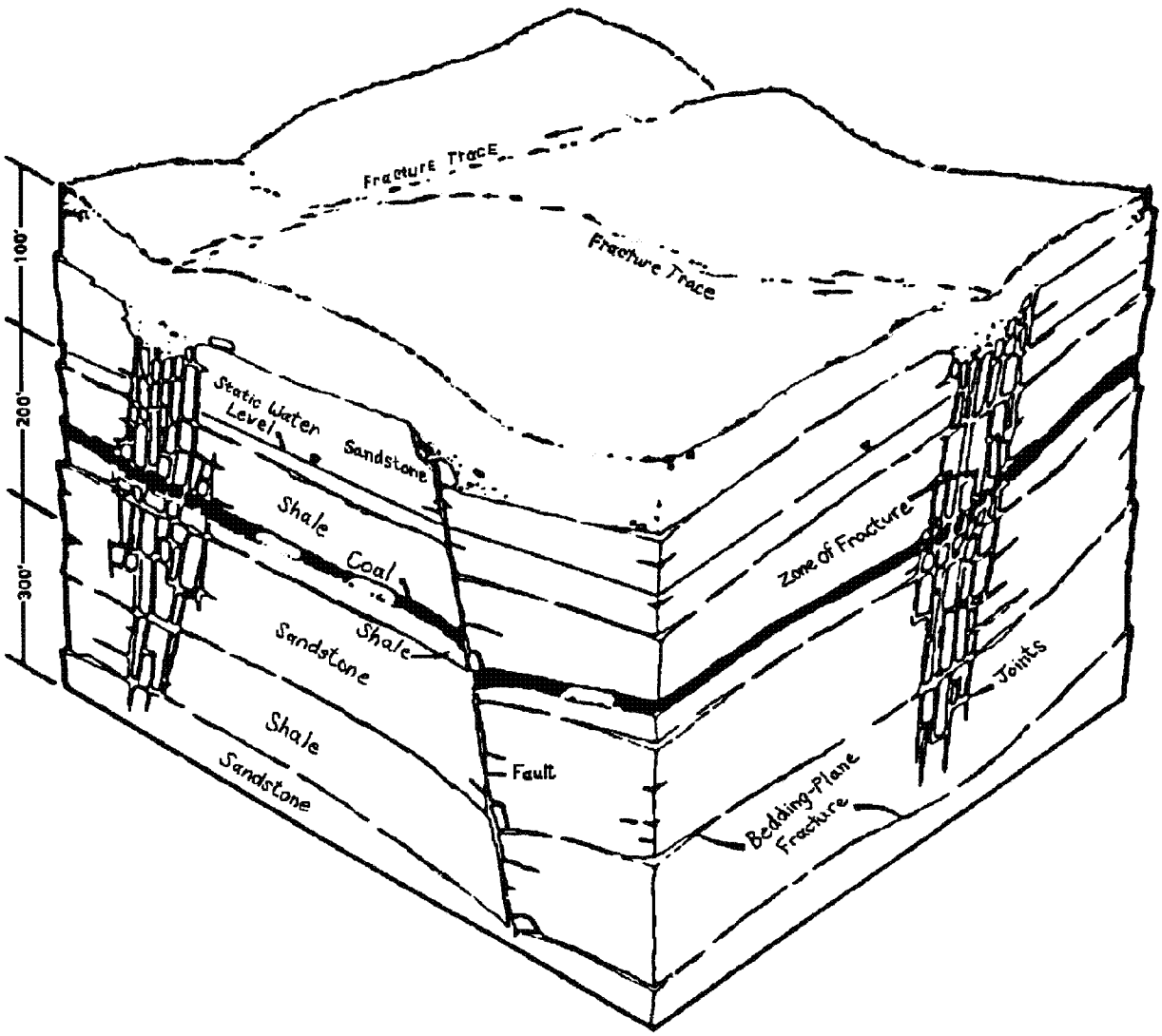


Figure 2.11 Idealized block diagram of fracture zones in the eastern coal region (modified from Ebaugh, 1986)

due to bending of rock units. Studies in Virginia regarding coal bed methane have documented highly permeable zones related to faults. Low-angle, imbricate thrust faults and broken rock were encountered in an Upper Freeport underground mine in western Pennsylvania, also yielding large quantities of gas and water when intersected (Elder et al.,1974; and McCulloch et al., 1975)

Geology

Pennsylvania's Appalachian Plateau physiographic province (Figure 2.13) extends from the western state border to the Allegheny Front. The Front is a prominent southeast-facing escarpment of approximately 1000 feet (305 m) of relief (Fenneman, 1938). The bituminous coal fields of Pennsylvania are almost entirely contained within this province. The effects of glaciers (extent shown on figure 2.13) on the northwest section of the plateau has served to scour and erode preglacial topography, fill topographic lows with glacial deposits thereby reducing relief, and in general produce characteristic glacial landforms such as drumlins and morainal features (Hornberger et al., 1981).

The rocks of the Appalachian Plateau are almost entirely sedimentary. Rock types consist of shale, sandstone, conglomerate, limestone, underclay, claystone, coal, and siltstone. The strata are relatively flat lying with northeast-southwest trending folds. Structural relief decreases north-westward in a step-like fashion from the well-defined folds of the southeastern

side of the Plateau where anticlines rise 800 to 2500 feet (244 to 1067 m) above adjacent synclines (Gwinn, 1964). The prominent structural forms of the coal bearing rocks of the Plateau, especially the anticlines of the Allegheny Mountain Section, may be reflected in the configuration of the present topographic surface except where deposition of the recent sediments, as in the Glaciated Section, has masked the underlying geologic structure (Hornberger et al., 1981).

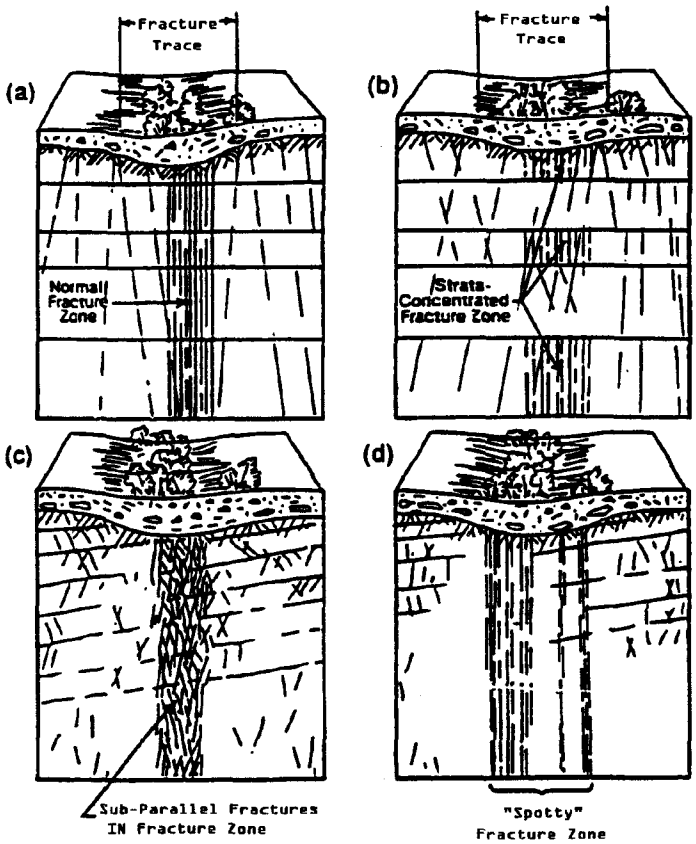


Figure 2.12 Styles of fracture development below fracture traces (Parizek, 1991)

Numerous factors combined to define the nature and extent of the cyclic deposits of sediments (cyclothems) which resulted in the repetitive coal measures of the basin. The result is a complex, somewhat cyclical, relatively flat-lying mix of sedimentary rocks with the prevailing characteristics being numerous lateral facies changes and interfingering lithologies.

Although referred to as cyclical, the sequence is generally not regular especially when considered in three dimensions. Marine zones, coals, and underclays are the most persistent units, with interfingering, pinching out, and local replacement by sandstone channels occurring commonly (Bieniawski and Mack, 1986). These thin, flat-lying, layered strata of contrasting permeability impart a layered heterogeneity, with a preferred horizontal flow component, upon the groundwater flow system. (See Chapter 8 for additional details regarding the region's geology.)

Aquifer "Types" of the Plateau

Since the rock matrix serves as a flow medium for groundwater, its origin and physical properties are controlling factors regarding the character of the flow system. Physical properties such as grain size, grain shape, degree of sorting, type and extent of cementation, degree of induration, and amount of compaction can affect the hydrologic properties of the rock mass. Sedimentary processes, which include weathering, transportation, deposition, and lithification, ultimately affect the resultant groundwater flow system by controlling certain of these physical properties. Shale, coal, siltstone, underclay, massive limestone, and tightly cemented sandstone generally display poor primary permeabilities (intergranular permeabilities) (Williams et al., 1982). Post-depositional processes can profoundly alter a unit's hydraulic conductivity through the development of secondary permeability features. On the Plateau, these secondary features govern the flow system's character.

Topographic relief affects head differences between areas of groundwater recharge and discharge, thereby influencing the depth of local groundwater circulation. In hilly terrain, such as is found in the northern Appalachian coal measures, the incised topography produces numerous subsystems within the major flow system. Water will often discharge to the nearest topographic low in this type of setting or it may continue to a more regional discharge area in the bottom of a major valley. Natural discharge occurs mainly as cropline springs and seeps, and base flow into streams and lakes. The water table located within the ridges between surface drainageways tends to become depressed to the level of the surface streams. However, semi-perched zones occur routinely within the ridges due to the numerous low permeability units.

Most porosity and permeability in the Appalachian Plateau is secondary. For example, bedrock aquifers usually include bedding-plane partings and near-vertical fractures, both secondary permeability features. Without such secondary features, the bedrock would likely not be a significant aquifer (Heath, 1984). However, the characteristics of the fractures (width, spacing, frequency, etc.) differ between, but may be consistent within, various stratigraphic units. Therefore, it is convenient to consider the stratigraphic units

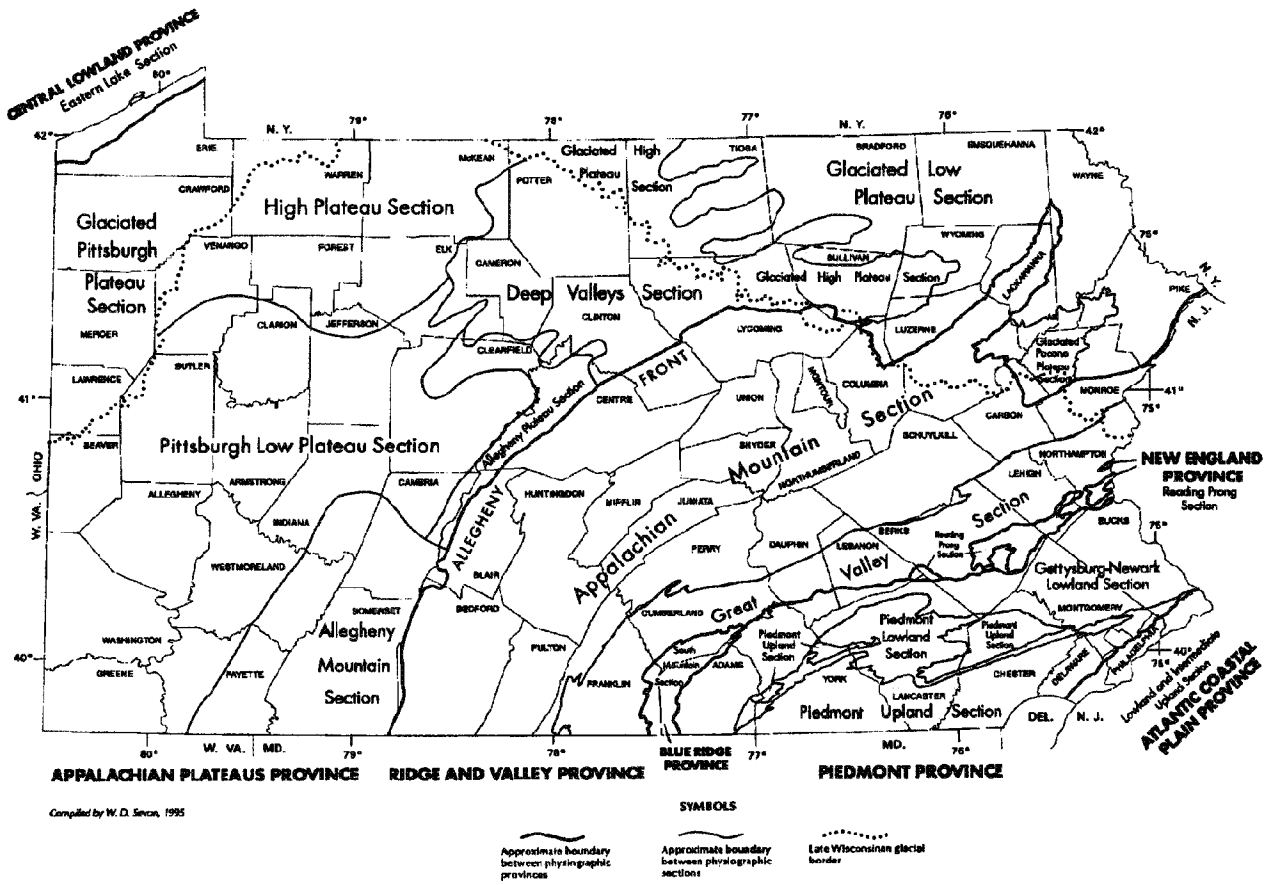


Figure 2.13 Physiographic Provinces of Pennsylvania (PA-DCNR MAP 13)

as the aquifers. The "typical" vertical profile on the Plateau consists of a shallow, unconfined (possibly including seasonally perched or semi-perched zones) system grading to a semi-confined system at intermediate depth.

Secondary openings such as joints, fractures, bedding-plane partings, zones of fracture concentration, fault zones, and slaty cleavage may greatly increase the water movement and storage within the coal measures (Parizek, 1979). Joint development within coals and associated strata add significantly to the vertical permeability of both aquifers and confining beds (Parizek, Sgambat, and Clar, 1979). Solution enlargement of secondary openings within marine and fresh-water limestones can transform these units of poor permeability into major aquifers (Parizek, White, and Langmuir, 1971). These features commonly counteract the previously outlined layered heterogeneity of the coal measures by augmenting vertical hydraulic conductivities allowing for significant vertical communication within the system. Additionally, rock types with relatively low primary permeabilities can become significant water producers due to interconnected fractures. Because of the importance of secondary permeability features to groundwater flow in this setting, the classic use of terms such as aquifer, aquitard, and confining layer can become obscured even on a local scale. A single lithologic unit, such as a shale, can be characterized as either an aquifer or an aquitard over a relatively short lateral distance depending on conditions such as proximity to the shallow weathered zone, proximity to stream valleys, and magnitude of open joints (generally decreasing with depth).

Equipotential lines are roughly parallel to bedding in low permeability zones and nearly perpendicular to bedding in high permeability zones. As a result, water flows laterally through high permeability units and vertically through low permeability units (Carswell and Bennett, 1963). Shales and other units with low permeability, still may transmit large amounts of water,

but transmit it vertically. A study in eastern Ohio estimated that 85% of the discharge from the local flow system was due to leakage through aquitards to deeper levels (Weiss and Razem, 1984). This contrasts with the findings that most groundwater in the stress-relief/weathered regolith subsystem never reaches the deeper flow systems. Weiss and Razem's study was in a low relief area, probably dominated by intermediate and/or regional flow systems. Nevertheless, their study suggests that low permeability units can transmit large amounts of water vertically. Most of the vertical change in head occurs across low permeability units (Carswell and Bennett, 1963; Peffer, 1991). The Appalachian Plateau contains substantial ranges of permeability. Permeability in sandstones may be hundreds of times greater than in shales (Brown and Parizek, 1971). Hawkins et al. (1996) also found permeability differences of over 100 times between the stress-relief/weathered regolith subsystem and the ridge-core subsystem. Schubert (1980), using laboratory-determined air permeabilities, showed the average ratio of horizontal to vertical hydraulic conductivities varied with lithology, ranging from 2.6 in sandstone to 14.9 in siltstone in rock samples from the coal fields of Pennsylvania.

Semi-perched aquifers occur when a relatively impermeable layer impedes the downward vertical flow of groundwater, but not to the extent that an unsaturated zone occurs uniformly beneath this flow-restricting zone. This type of aquifer is common on the Appalachian Plateau. They result where enhanced vertical hydraulic conductivity along the rind of a hill leads to step-like offsets in the water table near croplines. Under semi-perched conditions, the hill core remains saturated with perching occurring along the margins of the hill where the increased vertical permeability eliminates the "confining" properties of aquitard units resulting in free draining. The groundwater then migrates downward to a point where a sufficient permeability contrast exists to direct flow to a hillside spring (Figure 2.14).

Perched aquifers, where an unsaturated zone consistently exists beneath a poorly permeable zone, appear to be the exception rather than the rule on the Plateau. Their existence depends on consistent vertical permeability contrasts within the layered coal measures which is rare due to numerous and often abrupt lateral facies changes and extensive fracturing. By definition, a perched aquifer has unsaturated conditions both above and below (Figure 2.14). As a general rule on the Plateau, aquitard units within hills allow considerable leakage, thereby precluding the development of classic perched aquifers. Where they do exist, perched conditions are often temporary, dissipating with time due to downward leakage, sporadic recharge, and evaporation from the surface. Water chemistry data from the cores of hills versus cropline springs routinely indicate that the springs are fed by shallow dilute groundwater as opposed to water moving laterally from the middle of the hill. The chemistry data generally supports the semi-perched model where significant downward leakage occurs through aquitards in the cores of hills with only minor amounts of water moving laterally from hill cores to emanate as cropline springs.

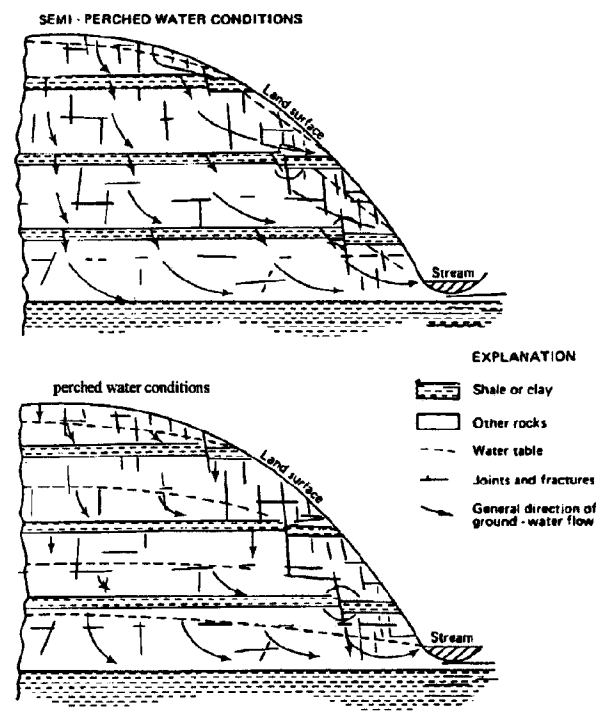


Figure 2.14 Idealized semi-perched and perched water conditions (modified from Ward and Wilmoth, 1968).

Confined aquifers are common at depth on the Appalachian Plateau. The majority of what are often termed confined aquifers, particularly those at intermediate depths, are better designated as semi-confined or "leaky" aquifers in that the associated aquitards have only marginal confining capabilities, i.e., are capable of transmitting significant quantities of water. Confining layers bounding aquifers need to be properly characterized particularly in cases where mine inflow may depend on the hydrologic properties of a nearby confining layer.

Many flowing wells in Pennsylvania are from confined aquifers that are recharged at their outcrops along the crests or flanks of anticlines (T. McElroy, personal communication, 1995). Of course flowing wells need not be from confined aquifers. In groundwater discharge areas, hydraulic head increases with depth and may eventually exceed the elevation of the land surface. A well in that type of setting, if cased to the appropriate depth, would flow. In Figure 2.10, a well drilled within the valley would flow at the surface, even though it is not tapping a confined aquifer.

Unconsolidated aquifers occur on the Appalachian Plateau in major alluvial valleys and terraces, and in glacial sands and gravels. Tills are usually capable of storing significant quantities of groundwater, but cannot transmit large amounts. In contrast to consolidated bedrock, primary porosity dominates in unconsolidated sediments. In certain areas such as the Allegheny-Ohio River valley of western Pennsylvania, alluvial deposits can be the most productive water-bearing formations. Alluvial valley aquifers receive most of their recharge from discharge from the underlying bedrock, and discharge into the rivers.

Groundwater Flow Systems

Recharge and Discharge Areas

Groundwater flow systems are commonly defined by their recharge and discharge areas. Groundwater routinely moves from recharge areas, generally topographic highs, to topographic lows. Recharge and discharge zones have unique "potential" signatures regardless of the flow system. In a recharge area hydraulic head decreases with depth causing downward flow; in a discharge area hydraulic head increases with depth causing upward flow (Saines, 1981). Flow systems can be classified as local, intermediate, or regional to characterize the areal extent of the flow system, the time required for groundwater to travel from the recharge to discharge areas, and the proximity of the discharge area to the most distant recharge area.

The relative size of the recharge versus discharge zones varies. However, for most topographic settings the hinge line is much closer to valley bottoms than to topographic highs. The "hinge line" is an imaginary line separating recharge areas from the discharge area. Discharge areas commonly constitute only 5-30% of the surface area of a watershed (Freeze and Cherry, 1987). Experience gained through locating and mislocating discharge zone piezometer clusters on the Plateau has indicated that upward gradients associated with stream valleys will be detected routinely only where the piezometers are in proximity to the stream, i.e. - the discharge area is very small compared to the recharge and lateral flow zones. Harlow and LeCain (1991), working in a similar setting in Virginia, confirm that upward gradients were detected only in proximity to streams.

Regions such as the Appalachian coal measures with their small basins, marked relief, and humid climate generally develop a groundwater flow system which can be readily broken into distinct parts - local, intermediate, and regional (Toth, 1963) (Figure 2.15). Superimposed on these systems (particularly on the shallow system) are additional distinct flow zones (subsystems) defined by the density, interconnectedness, and aperture of rock fractures. Although the ideal is to characterize the flow systems by defining the density, orientation, and transmissivity of fractures for each rock type and each distinct flow zone, Brown and Parizek, (1971) have demonstrated that useful groundwater flow models can be constructed with limited data by assuming uniform fracture distribution.

Local (Shallow) Groundwater Flow System

The shallow flow system underlies hills, discharges to local streams, and, to some extent, leaks downward into the deeper, intermediate system, which discharge to higher order streams at lower elevations (Duigon, 1987). In some areas, local systems include water which is "perched" above beds of lower permeability. This groundwater may then flow laterally due to permeability contrasts and discharge as springs above stream level.

Poth (1963) describes shallow groundwater in the Mercer, PA 15-minute quadrangle as circulating in a series of "hydrologic islands" (Figure 2.9). The dissected nature of the bedrock surface has resulted in hills, largely surrounded by valleys containing perennial streams. These hills constitute the hydrologic islands. This description can be extended over most of the Appalachian Plateau. A shallow, local groundwater flow system operates within each hydrologic island and is hydrologically segregated from the local groundwater flow systems in adjacent islands. The base of the local flow system (particularly for islands adjacent to first and second order streams) is a distance below the level of the stream valleys bordering the island, defined by the maximum depth at which groundwater originating within the hydrologic island will flow upward to discharge in the adjacent stream valley (see Figure 2.9).

Recharge to the local system is completely from within the hydrologic island. Discharge from the local system is into the adjacent stream valleys and via leakage into deeper intermediate and regional groundwater flow systems. In areas adjacent to larger streams and rivers, local groundwater which leaks downward may commingle with intermediate or even regional flow which is rising to discharge within the valley.

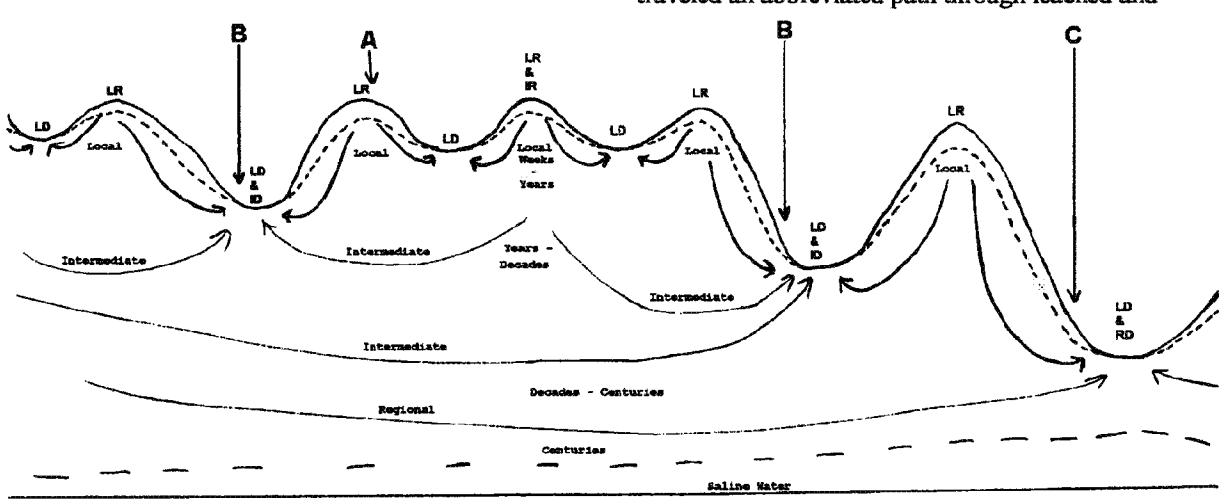
This shallow flow system is the area of the most active groundwater circulation. It is the zone which contributes water to the vast majority of domestic wells. Because hydraulic conductivity routinely decreases with increasing depth (studies have indicated an order of magnitude decline for every 100 feet (30 meters)), it is estimated that as much as 99.5% of total groundwater circulation occurs within the shallow (within 175 feet (53 m) of the surface) zone (Stoner et al., 1987).

The local flow system can often be divided into subsystems (Kipp and Dinger, 1987). Subsystems include the stress-relief fracture (Ferguson, 1967, 1974) and weathered regolith zone and the ridge-core subsystem which is controlled by lithology and tectonic fractures (joints and faults).

Stress-relief/weathered regolith subsystem - The stress-relief/weathered regolith subsystem exhibits distinctive groundwater flow and chemical characteristics, relative to the ridge-core subsystem and deeper flow systems. This is due to the non-reactive (weathered) nature of the transmitting medium and the short residence time of groundwater in this system.

The weathered regolith is a highly transmissive zone consisting of soil, unconsolidated sediment (colluvial, glacial, alluvial, etc.), and weathered, highly fractured rock. It has been documented to a depth of approximately 10-20 meters (Hawkins et al., 1996; Gburek and Urban, 1990; Merin, 1992). Weathering has removed most soluble minerals, and groundwater flowing through this material picks up little mineral matter. Because of the open nature of the fractures within this zone, the groundwater "flow-through" time is short and this subsystem allows a significant portion of the recharge to short-cut to local discharge points. Various hydrologic tests have shown hydraulic conductivities within this zone to be one to two orders of magnitude greater than in zones which are only marginally deeper (Schubert, 1980).

Water chemistry within the stress-relief/weathered regolith subsystem should not be used to characterize groundwater under deeper cover which flows through unweathered rock. For example, rock units closer to the ridge center may contain groundwater with significant alkalinity due to circulation through unweathered calcareous strata. In comparison, an outcrop spring at the same stratigraphic interval shows little or no alkalinity because it is largely fed by groundwater which traveled an abbreviated path through leached and



LR = Local Flow System Recharge Area
LD = Local Flow System Discharge Area
IR = Intermediate Flow System Recharge Area
ID = Intermediate Flow System Discharge Area
RD = Regional Flow System Discharge Area

Figure 2.15 Sketch illustrating various characteristics of the composite fresh-water groundwater flow system.

weathered rock along the "rind" of the hill. (Refer to Chapter 9 for information regarding premining water chemistry.)

Recharge to the stress-relief/weathered regolith subsystem is through ridge-top and valley-wall fractures. Groundwater flows through the interconnected bedding-plane partings and fractures to springs flanking the hill sides (frequently located on coal outcrops) and into stream channels where the fractures are exposed. Much of the water that enters this shallow subsystem never penetrates to the nearby ridge-core subsystem, nor to deeper flow systems. Residence time is as short as days to a week (Hawkins et al., 1996). Water levels in wells in the stress-relief zone respond quickly to precipitation events (Kipp and Dinger, 1987; Hawkins, et al., 1996). Wyrick and Borchers (1981) determined that stress-relief fractures significantly affect the surface water hydrology in Appalachian Plateau valleys. Their study in the Black Fork valley in West Virginia showed that stream flow per square mile of drainage area increased 6 to 11 times downstream from the outcrop of stress-relief and bedding-plane fractures in the stream bed.

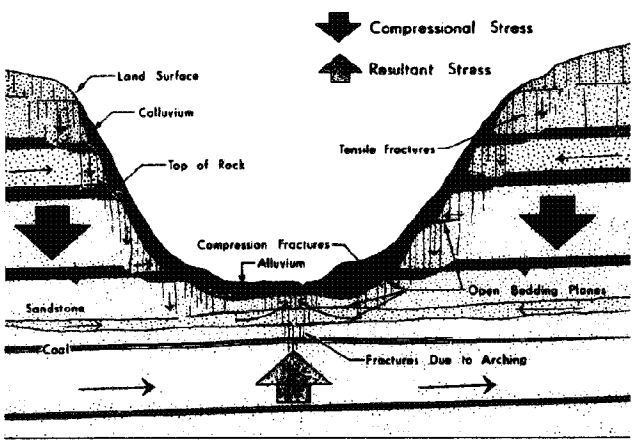


Figure 2.16 Generalized geologic section showing features of stress-relief fracturing and associated groundwater flow (modified from Ferguson, 1974).

Ridge-core subsystem - Flow within the ridge-core subsystem is controlled by lithology and regional joint sets. The ridge-core subsystem receives recharge through the stress-relief/weathered regolith subsystem. Groundwater flow is through tectonic fractures, bedding-plane partings, and to a much lesser degree, through intergranular porosity. Low permeability units (such as claystones and shales) exert more control within the ridge cores due to the lack of stress-relief joints and weathering which controls groundwater movement along the margins of the hills. Because the integrity of these low permeability layers has not been compromised fully in the ridge cores, as may be the case along the hillsides, groundwater can mound on these layers and either flow laterally to mix with groundwater within the stress-relief/weathered regolith subsystem and discharge to the local stream valley; or can leak downward to an intermediate or even regional flow system.

Residence times and response times to precipitation events within the ridge cores are intermediate between those for the stress-relief/weathered regolith subsystem and deeper systems. The ridge-core subsystem is part of the local flow system because it is part of the hydrologic island which discharges into the valley adjacent to the local recharge area.

Intermediate Flow Systems

Intermediate groundwater flow systems are those below the local shallow flow system but above the regional system (Richards, 1985). The intermediate system has some distinctive features which allow it to be separated out as a distinct zone. Although this zone may very well contain some components of the local and/or regional systems, the primary controls on groundwater flow are regional joint sets, bedding-plane partings, lithology, and zones of fracture concentration. At least one local flow system occurs between its surface recharge area and its discharge area. Recharge to intermediate systems is from leakage from overlying local systems, shallower intermediate systems, and at the basin divide of the defining recharge area. The flow passes beneath two or more hydrologic islands, similar to regional systems, but discharges in valleys above the lowest level of the drainage basin. Flow rates and residence times are generally between those of local and regional groundwater flow systems, probably varying from years to decades, depending on the level of the intermediate system and the length of the flow path. Vertically, at any given point, there is a single local flow system and a single regional flow system. However, there may be multiple intermediate flow systems at different levels, defined by multiple discharge areas, between the local and the regional flow systems (A in Figure 2.15).

Also, in intermediate flow systems regional structure tends to play a more important role in groundwater movement than with local flow systems. The important controls on the shallow system, such as incised topography with its associated undulating water table,

weathering, and fracture enhanced permeability, are less important at these depths.

Regional Flow Systems

A deep, regional groundwater flow system, which lies beneath the level of the low order stream valleys bordering the hydrologic islands and intermediate flow systems, operates independently of the shallower systems. The base of the regional system is the fresh water/saline water contact. The vast majority of groundwater circulation is primarily at shallow to moderate depths (< 300 feet (90 m)). Recharge to the regional system is from major drainage basin divides and leakage from multiple shallower (local and intermediate) systems. Regional groundwater does not flow to the surface to discharge in the stream valleys bordering the hydrologic islands (unless they are master streams), but continues beneath adjacent hydrologic islands and intermediate flow system discharge points to larger, deeper, regional discharge areas. These discharge areas are usually in larger, master stream valleys. These master stream valleys are commonly a major stream valley at the lowest level of the drainage basin (Richards, 1985). Within these regional groundwater systems flow rates are very slow and residence time is probably measured in decades or centuries.

Discussion

Local flow systems dominate in areas of high relief, while regional flow systems dominate in areas of low relief (Toth, 1963; Freeze and Cherry, 1987). Most surface mines are confined to a single hydrologic island. The majority of groundwater flow is probably in the local system, and most of that is probably within the stress-relief/weathered regolith subsystem (Harlow and LeCain, 1991). However, because substantial recharge to regional and intermediate flow systems consists of leakage from shallow (local) flow systems within the hydrologic islands, surface mines can potentially affect the deeper groundwater flow systems. Consideration of groundwater flow at a mine site requires recognition of at least the local and the first underlying groundwater flow systems because they are the most directly impacted by mining. Generally, little information exists for deeper intermediate and regional flow systems. Because a single hydrologic island comprises such a small part of the total recharge area for deeper flow systems, the effects of a single local flow system are probably insignificant to most of these deep systems.

There is little age dating of water on the Plateau. Wunsch (1993) used tritium (^3H) concentrations in nested piezometers on an unmined hill in the Eastern Kentucky coal field to obtain relative dates on the groundwater. Tritium dating is possible because high levels of tritium were introduced into the hydrologic cycle by large-scale atmospheric testing of nuclear weapons from 1953 through the 1960's. Pre-bomb tritium in precipitation was 5 to 20 tritium units (1 tritium unit = 1 tritium atom in 10^{18} H atoms). Post-bomb tritium levels were as high as thousands of tritium units. Tritium has a half-life of 12.3 years, so water with less than 2-4 tritium units is generally older than 1953 (Freeze and Cherry, 1979). The deepest piezometer installed by Wunsch was at the top of a hill and was open at a depth of 416 feet (126.8 m); approximately the same elevation as the adjacent second-order stream valley. The water at this depth, measured in 1991, had < 1 tritium unit. This suggests that this water is minimally decades, and possibly centuries, old. Water in the same nest at a depth of 195 feet (64 m) had a value of 2 tritium units, thus probably at least four decades old. Shallower water (168 feet (51.2 m) or less) in the hill core had values of 13 tritium units or higher, thus was conceivably younger than 1953. This tritium data provides some insight into groundwater ages for at least this area of the Plateau in Kentucky. The 416-foot (126.8-m) piezometer is probably in the intermediate flow system (possibly the deep ridge-core subsystem) and is decades or centuries old. Water in the shallow flow system (down to a depth of 168 feet (51.2 m) in the hill core) is probably several decades old or younger. The US Geological Survey has just begun to date Plateau water using chlorofluorocarbons, however this work has not yet been fully analyzed and has not been published.

Identification of Flow Systems

It is crucial to identify the various flow systems at a mine site. Several types of data will provide clues to delineating groundwater flow systems at a given location. These include physical, hydrochemical, and thermal data, discussed below, and the previously discussed age dating.

Physical data - Some physical data useful in obtaining a first estimate of the flow systems can be obtained from topographic maps. Hydrologic islands can be discerned on topographic maps to determine the approximate extent of the local flow system(s). Deeper flow systems can by estimated based on the location of probable discharge areas (large streams and rivers).

Boundaries between vertically adjacent flow systems (e.g., between the local and the first underlying system) may be determined from piezometer data, if available. The piezometric surfaces for aquifers 2 (in the local flow system) and 3 (in the first underlying flow system) on Figure 2.10 show how there can be abrupt head changes when passing from one flow system into another. Sufficient piezometer data to show flow direction might indicate flow in different directions in adjacent flow systems. Flow directions may diverge by 180° at flow system boundaries.

Discharge and water level fluctuation provides clues to the flow system contributing water to springs and wells. Shallow groundwater flow responds more quickly to precipitation. As a result, spring discharge and well water levels from the stress-relief/weathered regolith subsystem of the local flow system are often quite variable relative to springs and wells from the ridge-core subsystem or from deeper flow systems (Hawkins et al., 1996).

Hydrochemical data - Because of differences in rock mineralogy, residence time, and influence of the brine underlying the composite flow system, the chemistry of groundwater in different flow systems and subsystems varies.

Poth (1963) and Rose and Dresel (1990) identify three stages of "flushing" that roughly correspond with the three levels of the previously outlined flow systems. The deepest zone, directly affected by concentrated brines which exist at depth throughout all areas west of the Allegheny Front, is a NaCl-rich diluted brine zone. This zone is diluted with surface water, that has leaked from shallower flow systems, but retains appreciable amounts of both Na and Cl. This chemical signature is indicative of the more regional flow systems described above.

A shallower system (intermediate zone) exists in which Cl has been removed by flushing with surface waters, but considerable Na remains adsorbed to clays and similar materials, leading to the Na-HCO_3 waters that are commonly found at intermediate depths. The elevated Na is a result of cation exchange, with Na released from the exchange sites in response to replacement by Ca, Mg, and possibly Fe (especially with mine waters). Piper (1933) outlined this process as follows, "...Many of the water-bearing beds - whether they are sandstone, shale, or limestone - contain soft sodium bicarbonate water where they lie at intermediate depths. This soft water is believed by the writer to represent calcium bicarbonate water that has exchanged its calcium and magnesium for sodium by reaction with base-exchange silicates in the rock as it has percolated downward along the dip of the water-bearing bed. The hardness due to the bicarbonate of calcium and magnesium is removed in proportion to the completeness of the exchange reaction, and the water finally passes into the sodium bicarbonate type..."

In the upper-most zone Na is completely flushed leaving a Ca-HCO_3 water typical of shallow groundwater. The shallow flow system is further divided (Brady et al., 1996) into a low dissolved-solids zone associated with the stress-relief/weathered regolith subsystem, and a zone with higher dissolved solids associated with unweathered rock (ridge cores). (This is discussed in greater detail in Chapter 9).

Because stream valleys function as sumps for the discharge of both fresh and saline groundwater, the contact between the fresh and saline groundwater probably "cones up" beneath the streams and lies at successively greater depths away from the streams. Mining activity and practices for controlling water quality or quantity may alter the depth to saline water and the amount of discharge of saline water to the streams (Hobba, 1987).

Wunsch (1993) developed a conceptual "hydrochemical-facies" model (Figure 2.17) for an unmined ridge based on site-specific data from eastern Kentucky. According to Wunsch (1993, p. 72), ..."The model shows four zones where the major cations and anions comprising the water type for a particular water sample could be predicted with a high degree of probability. The model shows a depressed salt-water interface below the ridge due to downward movement and accumulation of fresh water. The hydrostatic pressure imparted on the salt water is transmitted in the salt-water zone, causing it to rise at locations where fractures breech confining layers (valley bottoms)..."

Data collected at numerous mining sites in southwestern Pennsylvania (including the 580 Pocket site which is incorporated as Case Study No. 1 at the end of this chapter) support Wunsch's hydrochemical model for shallow groundwater flow on the dissected Appalachian Plateau of Pennsylvania.

Thermal Data - Groundwater temperature, relative to annual mean air temperature, and annual fluctuations of groundwater temperature can provide clues to help delineate the flow systems. Groundwater flow

serves to redistribute terrestrial heat. "In the upper 10 m or so, diurnal and seasonal variations in air temperature create a zone that is thermally transient."(Freeze and Cherry, 1987). Within the local flow system and especially within the stress-relief/weathered regolith subsystem, there is usually an annual groundwater temperature fluctuation of up to 10° C. The mean annual groundwater temperature within shallow groundwater (at the base of the zone of groundwater temperature fluctuation) is generally 1° to 2° C higher than the mean annual air temperature at that site (Heath, 1989). Beneath the zone of groundwater temperature fluctuation, the temperature increases with depth at a rate dependent on the geothermal gradient. Van Orstrand (1918), working in Washington County, PA, measured groundwater temperatures at a gas well site to a depth of 7000 feet (2296.6 m). Between depths of 100 to 2000 feet (32.8 to 656.2 m) he found temperature increasing at a rate of 1°F for each 110 feet (36 m) of depth). According to Piper (1933), the average groundwater temperatures observed in wells in southwestern Pennsylvania were; for wells less than 100 feet (32.8 m), 51.8°F: for wells between 100-200 feet (32.8-65.6 m), 52.1°F; and for wells between 200-450 feet (65.6-147.6 m), 52.5°F.

Discussion - The above data types will often be helpful in determining whether a particular groundwater sample is from the stress-relief/weathered regolith subsystem, the ridge-core subsystem, or from a deeper flow system. Generally, the shallowest flow will be low in dissolved solids, variable in flow and temperature, and discharge along the flank of a hydrologic island. Springs in intermediate and regional discharge areas may show characteristics of local groundwater flow (B and C on Figure 2.16), such as the example in Table 2.1 for a proposed mine in the Yellow Creek valley in Indiana County. Springs SG-10 and SG-15 are higher in pH, specific conductance, and alkalinity than other springs, and are similar in those parameters to water from nearby wells. These springs probably produce water from an intermediate flow system discharging into the Yellow Creek valley. Other springs produce water that originated as recharge on the adjacent uplands, and are discharges of the local flow system. (Another example of springs representing both local and intermediate flow systems is discussed in Chapter 9.)

Table 2.1

sample pt.	Median pH	Mean Specific Conduct.	Mean alk.(mg/l)
springs 10 and 15	7.59	428	105
wells	7.56	595	170
other springs	6.5	144	9

Case Studies

Case Study No. 1 - 580 Pocket Mine

This underground coal mine complex is located in Greene and Susquehanna Townships, Indiana County, PA. The mine is located on the Appalachian Plateau in an area of broad, rolling uplands that is drained by incised, dendritic stream valleys typically 200 to 500 feet

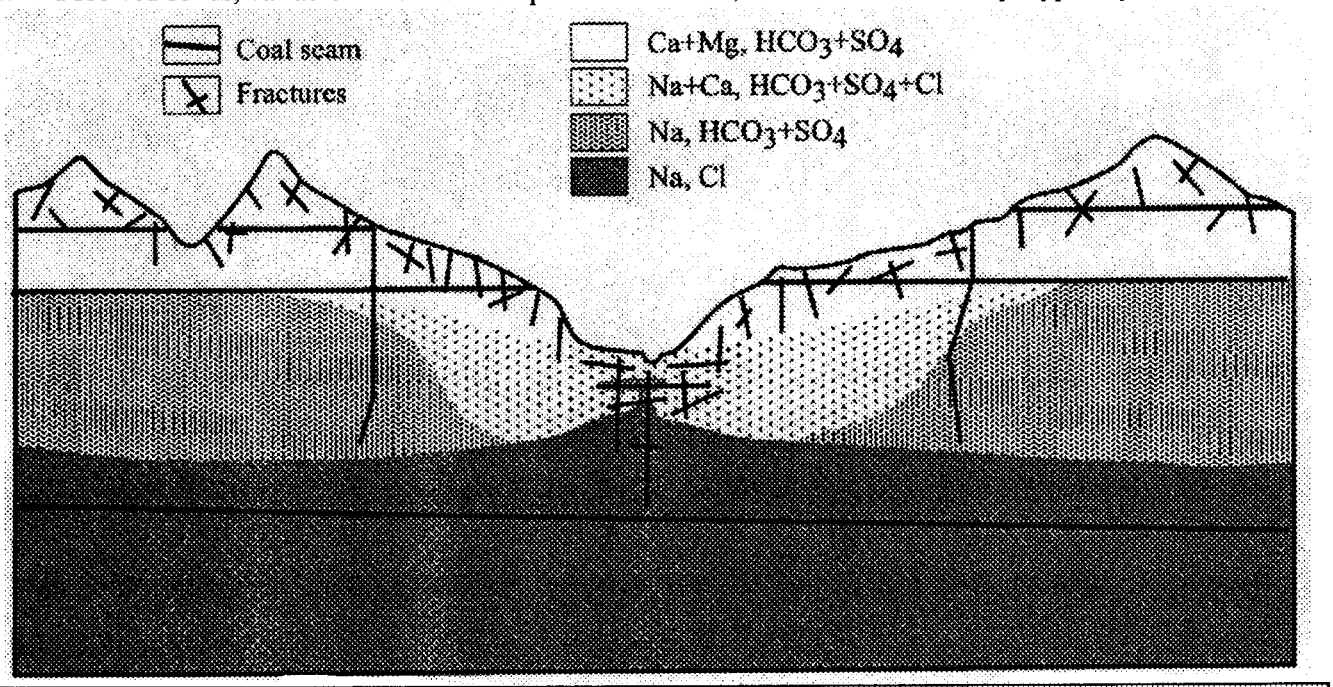


Figure 2.17 Conceptural model of hydrochemical facies (Wunsch, 1995).

(60 to 150 m) deep. The geologic structure is essentially flat-lying sedimentary rocks which have been folded such that the folds generally dip at two to six percent to the west toward the Brush Valley syncline. Rock types consist of shale, sandstone, siltstone, claystone, limestone, coal, and underclay. The overburden sequence includes the upper portion of the Freeport Formation (Allegheny Group) and the lower two-thirds of the Glenshaw Formation (Conemaugh Group) (see Figure 2.18).

The 580 Pocket is part of a large underground mine complex consisting of the approximately 3500-acre (1420-hectare) North Mine and the approximately 4550-acre (1840-hectare) South Mine which includes the 580 Pocket (see Figure 2.19). The North and South Mines are physically separated by a 250-foot (76-meter) wide unmined zone. The mine complex, particularly the South Mine and 580 Pocket, has been studied and evaluated over the past decade due to mining-induced dewatering episodes at domestic wells and superjacent streams. Additional mining related problems developed after the cessation of mining in 1992 when the pumps were turned off and the mine began to flood.

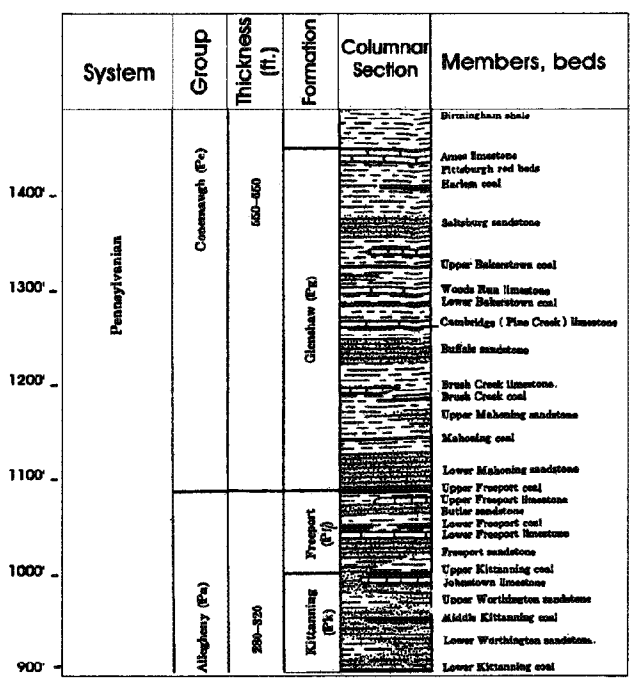


Figure 2.18 Generalized geologic column, 580 Pocket area.

The mine pool eventually emerged as seeps in the valleys above and adjacent to the 580 Pocket section of the South Mine.

Mining of the complex began in the Lower Freeport seam in 1969 and progressed down dip toward the Brush Valley synclinal axis (see Figure 2.19). Longwall, room and pillar, and room and pillar with retreat mining were all conducted within the complex. Typical cover ranged between 250 and 450 feet (76 to 137 m). Mining height was typically four to six feet (1.22 to 1.83 m).

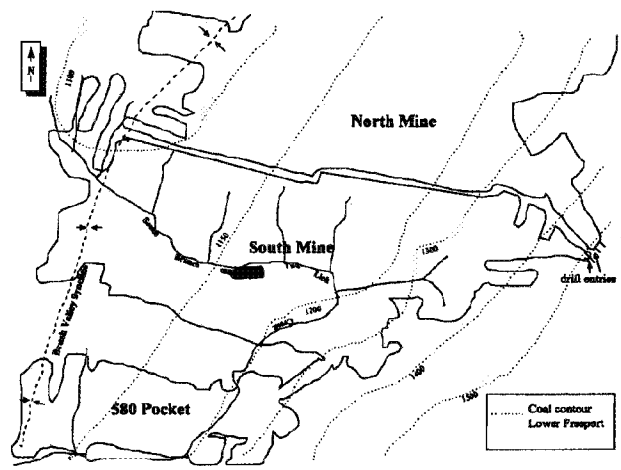


Figure 2.19 Map showing relative positions of sections of Greenwich mine complex.

Mining in the South Mine started at a drift opening (see Figure 2.19). Inconsistent seam thickness caused the mining company to leave a section south of the South Mine unmined. However, mineable coal was present farther to the south and a series of entries was driven through the low coal zone to access the 580 Pocket.

Due to the stream dewatering problems which accompanied the mining (1985 through 1994) and the subsequent postmining pool breakout in 1994, the South Mine has been and remains one of the most scrutinized and best instrumented underground mine sites in Pennsylvania. More than 50 piezometer points (many including continuous recorders) have been developed over the past decade at the site. Extensive surface and groundwater quality sampling has been conducted by various consultants and by Pennsylvania DEP staff. Additional site work has included: geophysical logging, bedrock coring, aquatic surveys, slug tests, pumping tests, whole-rock geochemical testing, and petrographic analysis.

Site Characteristics - Beginning in 1985, reports of stream diminution were being received by the Pennsylvania Fish Commission for sections of the South Branch of Two Lick Creek above the South Mine workings. High extraction mining, including a series of longwall panels, was conducted throughout the South Branch of Two Lick Creek basin during the 1980's and early 1990's. Dewatering episodes in the upper portions of the stream precipitated a number of hydrologic studies which were conducted within the basin beginning in 1988. Weirs and piezometer nests were installed in 1988 within the 580 Pocket area to characterize stream and groundwater flow and assess mining impacts. Impacts from mining in terms of groundwater level declines and stream base flow reductions were documented. Subsequent approvals for additional mining within the 580 Pocket section of the South Mine were conditioned to limit mining below and adjacent to the South Branch of Two Lick Creek and Repine Run watersheds. Mining was restricted to "first mining only", an extraction rate of approximately 55%, beneath stream valleys. This was an attempt to protect the perennial nature of the South Branch's lower sections and reduce impacts to Repine Run by minimizing the likelihood of subsidence fractures which may capture stream flow. This limited mining continued within the 580 Pocket until March of 1992 when mining ceased in the South Mine and the workings began to flood.

During 1992 and 1993, as the mine flooded and groundwater levels recovered, an ongoing hydrogeologic evaluation was conducted to document conditions during mine closure. The evaluation included stream gauging, mine pool level monitoring, and water quality sampling. Due to the dewatering which accompanied the mining there were concerns regarding development of postmining seeps along the impacted stream valleys below the final pool elevation. In February of 1994, contemporaneous with the mine pool potential reaching the stream bottom elevation, mine pool breakouts occurred to the South Branch of Two Lick Creek and Repine Run at several locations near their confluence. The seeps bore the standard high sulfate, high specific conductance signature of mine water. Additionally the seeps exhibited a high sodium signature which is often indicative of the intermediate flow system on the Plateau where considerable sodium remains adsorbed to clays due to cation exchange as outlined earlier in this chapter. This sodium signature is therefore an indicator of water from a deeper part of the flow regime as opposed to the usual Ca-HCO_3 signature water which is common for the shallow flow system. The mine pool also exhibited this sodium marker.

Immediately following the discovery of the seeps a more comprehensive hydrogeologic site characterization was conducted. Eight additional piezometer nests were installed, rock coring from the surface to the Lower Freeport seam was conducted, and geophysical logging was performed at all drill sites. Chemical sampling of the mine pool and the surface seeps was initiated and included major anions and cations. Adjacent mine discharges were inventoried and sampled. Weirs were constructed at seeps with measurable flows. Additional aquatic surveys were conducted to evaluate the impacts of the discharges on the aquatic community and related stream uses. Additional testing included slug tests, pumping tests, and petrographic analysis.

Figure 2.20 (located in the pocket at the rear of this book) is a flow net depicting conditions at the down-structure end of the mine. The flow net is based on measured piezometric head values (after mine pool stabilization) at the site. This flow net is a semi-qualitative, conceptual model of groundwater flow at the site. The ridges between the valleys act as groundwater recharge areas. Water entering the ridge tops discharges locally to nearby streams, exits at contact springs, or continues downward to recharge deeper portions of the flow system. As can be seen, the arrangement of equipotential and flow lines has been predictably altered at depth due to the presence of the mine-void aquifer. The mine-void aquifer is highly transmissive and has essentially uniform head across the mined area. This is an important factor since areas within and adjacent to the down-structure portions of the mine workings will realize abnormally high post-mining heads relative to their premining head potentials. Where the mine void meets virgin coal, the head will reestablish the necessary gradient to drive water down structure through the intact coal aquifer. However, in stream valleys at the margins of the mine workings a large reservoir of mine water exists with head potential exceeding the surface elevation.

As pointed out previously in this chapter, stream valleys are typically locations containing a disproportionate number of secondary permeability features. This combination of conditions (head potential at depth exceeding the surface elevation; a large reservoir of highly ionized mine water; and fractured, transmissive overburden within the valley settings along the margins

of the mine workings) resulted in the postmining mine pool discharges to the South Branch of Two Lick Creek and Repine Run.

When considered against the backdrop of the previously outlined conceptual groundwater flow model for the northern Appalachian Plateau, the mine water emerged at the most logical location. The following is a listing of site-specific data and findings and their relationship to significant, ubiquitous features of the conceptual groundwater flow model outlined previously in this chapter. Information gathered during the hydrogeologic studies conducted at the 580 Pocket mine site is in accord with, and reinforces, the established groundwater flow model for the Appalachian Plateau of Pennsylvania.

Shallow flow system

Weathered surface zone - Data collected during the 580 Pocket site characterization confirmed the presence of a shallow, weathered, highly tranmissive zone devoid of significant, readily leachable or oxidizable minerals. Representative water quality data from shallow piezometers (in zones not affected by upwelling mine water) and from hillside springs are compiled in Table 2.2.

Water data (Table 2.2) are consistent with what would be expected within the shallow flow zone along the rind of the hill.

Site drill logs typically show weathered, fractured, and broken material down to a depth of about 15 meters (40-50 feet). This shallow zone routinely bottoms out at about 18-20 meters (55-65 feet) as documented by extensive on-site drilling.

Stress-relief fractures - At the 580 Pocket site the postmining seepage was valley associated. The valleys above and adjacent to the mine workings served as natural pressure release zones where the mine pool could migrate to the surface along valley-related fracture sets (valley stress-relief fractures and zones of fracture concentration). In addition to valleys being proximal to the discharges, other evidence indicates the presence of valley stress-relief fractures and valley-related fracture sets at this site, including: (1) valley corings showing a consistent fracture profile; (2) piezometer pumping tests demonstrating vertical communication; (3) mining-associated dewatering events indicating direct hydrologic communication between the shallow flow system and the mine; (4) the similar groundwater chemical signature throughout the vertical section between the mine level and the stream valley floor; (5) fracture trace analyses indicating that certain straight stream segments are fracture related; and (6) on-site research which found that fracture densities and orientations were statistically relevant factors to consider when evaluating conditions contributory to unstable mine roof. Each piece of evidence is addressed below:

(1) rock corings from valleys:

Broken and fractured rock is indicated in core logs to a depth of 150-200 feet (46-61 m) in valley settings at the 580 Pocket site. Core 95-08 included a rock quality designation (RQD) run contemporaneous with drilling. The RQD data indicate a steady progression from poor to excellent quality rock at a depth of approximately 140 feet (42 m). The RQD data for Core 95-06 shows a similar pattern - i.e., rock quality improving with depth to approximately 190 feet (52 m). RQD is based on the ratio of recovered core length to the total length of the core run, or to a defined section of the core. Only recovered core pieces four inches or longer are counted. In consolidated rock, it can be a good indicator of the degree of fracturing within the rock mass. A high RQD points to infrequent fracturing and a low RQD points to more frequent fracturing of the rock mass.

(2) pumping exercises at piezometer clusters:

A five-day pumping test was conducted at piezometer cluster 95-04 in late September-early October, 1995 as part of the site evaluation. The 95-04 piezometer cluster is open to six different horizons. These monitoring zones are spaced over the 266-foot (81.08-meter) distance between the mine level and the surface (see Figure 2.21). The 198-foot (60.35-meter) piezometer was pumped while monitoring was conducted (with pressure transducers) at the other five intervals. All monitored intervals showed water level declines and recoveries concurrent with the pumping event.

A similar pumping test was conducted on piezometer 95-07. Figure 2.22 shows reactions within the vertically segregated monitoring zones of piezometer cluster 95-07.

(3) dewatering events during mining:

As pointed out earlier in this chapter, dewatering of surface streams above the Greenwich mine workings was occurring for approximately seven years (1985 to 1992) prior to mine closure. A study conducted in 1988 and 1989 by a Greenwich Collieries consulting hydrogeologist documented mining-induced dewatering

Table 2.2 Typical shallow (weathered) zone groundwater chemistry (mg/L)

sam. pt	depth	date	alk.	SO_4	Ca	Mg	Na	sp.cond. (umhos/cm)
piez. 9506	35 ft.	10/5/95	38	36	38	7	13	88
"	"	10/24/95	12	21	4	1	6	62
"	"	12/13/95	12	19	5	2	4	62
"	"	2/27/96	6	45	6	2	4	72
DM-3	spring	3/6/96	12	11	-----	-----	3	136
TS-8	spring	3/6/96	7	17	-----	-----	2	62
TS-7	spring	3/6/96	2	20	-----	-----	2	56
TS-6	spring	3/6/96	2	19	-----	-----	2	58
TS-5	spring	3/6/96	5	17	-----	-----	2	61
TS-4	spring	3/6/96	4	19	-----	-----	2	60
ST-101	spring	3/6/96	2	17	-----	-----	2	54
ST-102	spring	3/6/96	2	15	-----	-----	2	50
RPD3/6-1	spring	3/6/96	5	14	-----	-----	3	52
ST-123	spring	3/13/96	9	24	-----	-----	2	74
RPD3/6-2	spring	3/6/96	12	15	-----	-----	2	75
RPD3/6-3	spring	3/6/96	11	16	-----	-----	2	86
RPD3/6-4	spring	3/6/96	7	14	-----	-----	3	84
RPD3/6-5	spring	3/6/96	2	14	-----	-----	2	58
RPD3/6-6	spring	3/6/96	2	18	-----	-----	2	62
RPD3/6-7	spring	3/6/96	2	19	-----	-----	2	65
RPD3/6-8	spring	3/6/96	2	19	-----	-----	3	63
RPD3/6-9	spring	3/6/96	2	17	-----	-----	2	64

in the form of reduced base flow to Repine Run. The study relied on stream flow measurements upstream and downstream of mining, groundwater level monitoring, and groundwater and surface water "control" points outside the influence of the mining to define baseline conditions.

(4) groundwater chemical signatures:

95-07 is the only piezometer cluster located along the center line of a valley. The piezometers within the 95-07 cluster are spaced at depths of 37 feet (11.28 m), 59 feet (17.98 m), 104 feet (31.7 m), 153 feet (46.63 m), 190 feet (57.91 m), and 235 feet (71.63 m)(coal seam). Chemical data are listed in Table 2.3. The piezometers are located in a groundwater discharge zone. The water chemistry has a similar signature along the entire vertical extent of the cluster.

(5) fracture trace analysis:

A consultant for the mine operator performed a fracture trace analysis based on air photograph interpretation. Several stream valley segments line up with the mapped fracture traces. They include: (1) portions of the South Branch of Two Lick Creek above the confluence with Repine Run; (2) Rock Run; and (3) several unnamed tributaries to South Branch of Two Lick Creek. A significant shortcoming of this mapping exercise is the lack of any ensuing field verification.

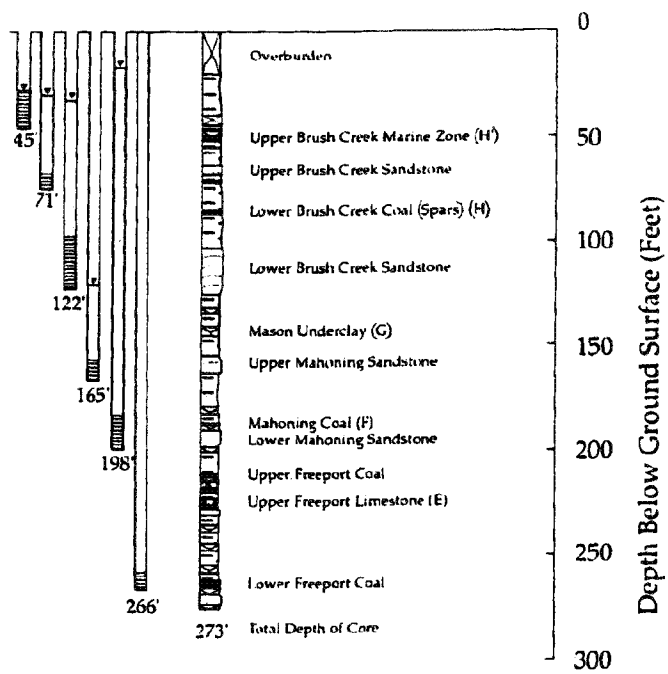


Figure 2.21 Piezometer 95-04 schematic

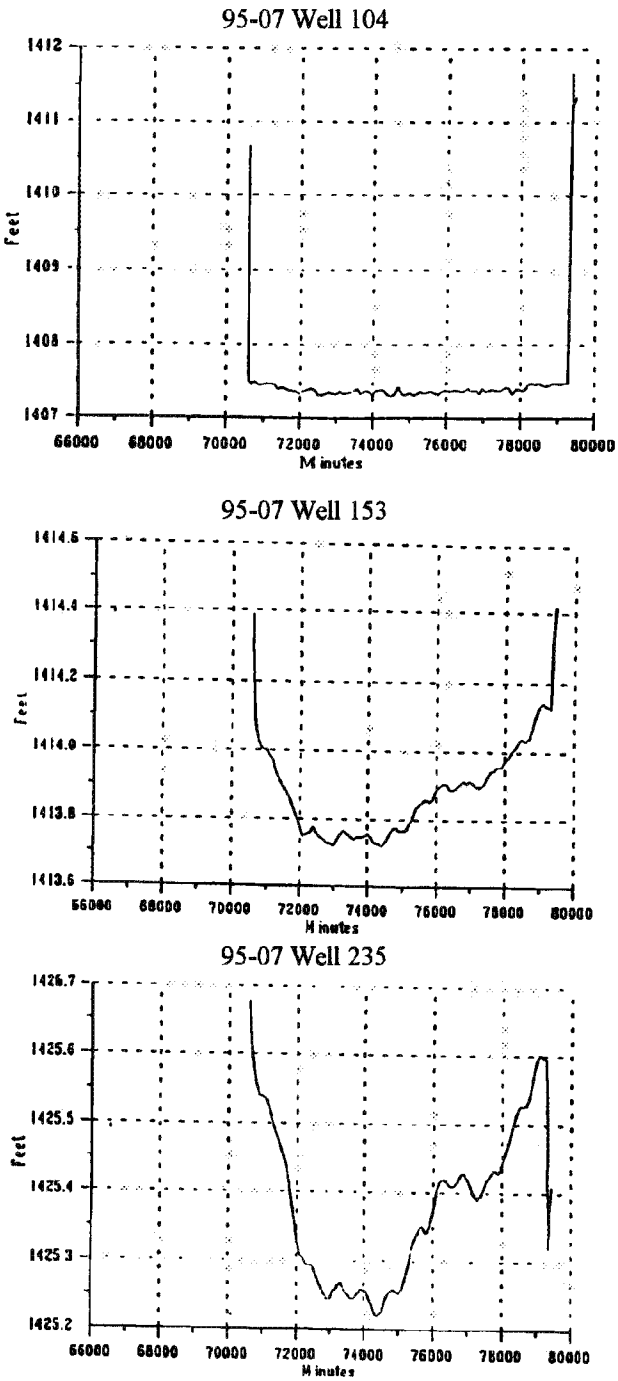


Figure 2.22 Groundwater responses at 95-07 piezometer cluster during pumping interval (pumping rate, 11 gpm).

(6) mine roof stability research:

Research conducted at the Greenwich mine in 1987 (Blackmer, 1987), with the goal of developing valid parameters for a Roof Stability Index (used to predict general mine roof stability), focused on "controls" which are significant when assessing mine roof stability. The four statistically relevant controls were geologic structure, fracture trace orientation, fracture trace density, and density of clay veins. This work points out the significance of naturally occurring fracture sets at mine depth.

Table 2.3 Water chemistry for piezometer cluster 95-07 (mg/L)

piez.	date	alk	spec. con. (umhos/cm)	Na	Ca	Mg	SO_4
37 ft.	2/22/96	504	1550	252	137	18	471
59 ft.	10/25/95	601	2050	570	63	13	561
59 ft.	2/22/96	560	1700	370	95	14	459
104 ft.	2/22/96	607	2000	420	76	15	564
153 ft.	10/10/95	616	2100	625	56	14	646
153 ft.	10/25/95	610	2050	660	35	8	550
153 ft.	2/22/96	611	2050	470	51	10	575
190 ft.	2/22/96	560	1800	420	46	11	453
235 ft.	10/25/95	23	1220	370	3	0.3	440
235 ft.	2/22/96	142	1240	250	2	0.1	450

Ridge cores - Groundwater samples pulled from piezometers located in ridge cores between stream valleys show a markedly different chemical signature than nearby spring data and data from shallow, weathered-zone piezometers. Table 2.4 lists the typical chemistry for the ridge cores at the Greenwich Mine. These points are located above the final mine pool level or in recharge areas with a strong downward flow component precluding the upward migration of mine water. The water is a Ca-HCO_3 type with low sulfate concentrations and low sodium concentrations.

Intermediate Flow System - Some difficulty in delineating the intermediate flow system at the 580 Pocket site occurs because of chemical and hydrologic influences from the deep mine complex. Even the sodium signature, which is an indicator of intermediate zone chemistry, can be influenced indirectly by the mining. For example, iron in the mine water may bump sodium ions from their exchange sites thus further elevating sodium levels. Additionally, the characteristic flow patterns of an intermediate flow regime, e.g., bypassing of local discharge points, can be mimicked by the insertion of a highly transmissive mine-void aquifer at depth. No piezometers were developed below the Lower Freeport coal horizon. Therefore comparisons cannot be made between the mine pool chemistry and deeper groundwater. Another limiting factor is the that chemical data prior to mining and flooding did not include analyses for major anions and cations (e.g., Na).

Table 2.4 Ridge core groundwater chemistry (mg/L)

piez.	depth	date	Alk.	spec. cond. (umhos/cm)	SO_4	Ca	Mg	Na
95-03	88 ft.	8/21/95	146	435	55	70	12	----
95-03	88 ft.	9/13/95	160	470	46	69	11	17
95-03	88 ft.	10/13/95	150	470	46	70		14
95-03	88 ft.	10/23/95	155	450	88	69	10	13
95-03	88 ft.	12/13/95	80	350	60	41	10	7
95-03	88 ft.	2/28/96	90	300	39	40	9	2
95-01*	181 ft.		134-199		41-75		----	------

* Data for piezometer 95-01 is presented as a range of values

Regardless of mine influences, the presence of sodium at mine depth indicates this zone is in the intermediate system. Table 2.5 lists representative mine pool and piezometer samples of this deeper, intermediate system. In addition to high concentrations of Na, SO_4 is also elevated. The SO_4 results from pyrite oxidation which occurred as a consequence of mining.

580 Pocket Site Versus Conceptual Model - The 580 Pocket site-characterization studies are consistent with the outlined conceptual flow model of the Appalachian Plateau. One important aspect of fracture flow is that the significance of any given fracture is scale dependent (major fractures on one scale can become minor fractures on another scale). Therefore, even for areas where fracture flow is dominant, sites can be modeled based on continuum assumptions if the fractures are relatively consistently spaced, are routinely interconnected, and the scale of the study area is large enough to treat as porous media. The fracture controlled nature of the groundwater flow system at the 580 Pocket site appears to manifest itself in the direct connection between the mine and the stream bottom and in the discrete, point source character of the seeps. Regardless of the controls on the flow system (secondary permeability features or intergranular flow), the seeps would have in all probability manifested themselves in the same location. The overriding controls on the occurrence and location of the seeps are: (1) the introduction of an highly transmissive, man-made aquifer at a relatively shallow depth; and (2) the related permeability contrast between the man-made aquifer and the adjacent rock mass which controls the mine pool elevation.

Table 2.5 Intermediate zone groundwater chemistry (mg/L)

samp. pt.	Depth	date	spec. conductance (umhos/cm)	alkalinity	sulfate	sodium	calcium
T-8 BH	mine pool	9/12/95	2200	639	670	530	58
T-8 BH	mine pool	10/2/95	2300	614	646	535	82
T-8 BH	mine pool	10/23/95	2250	640	713	635	39
T-8 BH	mine pool	12/12/95	2100	640	574	490	46
580 shaft	mine pool	2/21/96	2300	550	833	360	202
580 shaft	mine pool	2/22/96	2200	540	1000	430	176
95-08	212 ft.	11/21/95	2000	609	583	-----	49
95-08	212 ft.	12/13/95	2100	625	574	485	64
95-08	212 ft.	2/15/96	2050	610	559	520	72
95-06	272 ft.	10/5/95	3500	861	286	191	207
95-06	272 ft.	10/25/95	2950	685	130	288	4
95-06	272 ft.	12/13/95	1300	435	119	265	2
95-06	272 ft.	2/27/96	1260	450	108	280	3
95-07	235 ft.	10/25/95	1220	23	440	370	2.5
95-07	235 ft.	2/22/96	1240	142	449	250	2.1
95-04	266 ft.	9/12/95	4500	8	2850	945	246
95-04	266 ft.	10/13/95	4000	78	2215	840	234
95-04	266 ft.	10/24/95	3400	90	2040	840	191
95-04	266 ft.	12/13/95	3000	75	1630	520	158
95-04	266 ft.	2/27/96	3100	180	1705	690	183
95-03	374 ft.	8/21/95	1380	640	645	------	28
95-03	374 ft.	9/13/95	2350	652	615	665	27
95-03	374 ft.	10/12/95	2300	619	590	600	31
95-03	374 ft.	10/23/95	2300	617	648	745	32
95-03	374 ft.	12/13/95	2300	625	583	550	26
95-03	374 ft.	2/28/96	2000	690	538	500	24

Case Study No. 2 - Kauffman Mine

The Kauffman mine is a bituminous surface mine located in Boggs Township, Clearfield County, Pennsylvania. The topographic setting is fairly typical of the bituminous coal mining region of Pennsylvania's Appalachian Plateau - i.e., broad uplands dissected by deeply incised tributaries. The permit is bounded by a major state highway to the east (SR0153), Clearfield Creek to the west, and two northwest-flowing tributaries to Clearfield Creek, Camp Hope Run to the north and Sanbourn Run to the south. The topographic relief ranges from about 200 feet (60 m) near the headwaters of these tributaries to more than 525 feet (160 m) where the western hillside drops off to the level of Clearfield Creek. The soils of the hilltops are thin and sandy, reflecting the abundance of sandstone within the coal overburden. A cover of colluvial material blankets the relatively steep hillsides and clay-rich alluvium fills the valley bottoms adjacent to the streams (see Figure 2.23).

At the time this volume was written the Kauffman mine was being mined as a "demonstration" surface

mining permit. That is, the operator plans to demonstrate, by special handling high sulfur spoil and importing alkaline material, that they can mine this site without creating AMD. It should be noted that past mining at an adjacent mine (Thompson mine) on the same coal seam, just east of State Route 0153, resulted in severe AMD pollution to the surface water and groundwater. The Thompson mine has impacted both Sanbourn Run and Camp Hope Run to the extent that aquatic life is no longer present. Additionally, a pollutive groundwater plume of considerable lateral and vertical extent has migrated westward from the Thompson mine to beneath the eastern portion of the Kauffman permit. The complexities of this demonstration mining project, with respect to the site's geology, hydrogeology, and influences of past mining, necessitated the collection and study of a large volume of data. Many researchers, consultants, students, and other investigators have examined the site. Some of the published literature related to the Kauffman site includes Abate (1993), Evans (1994), Rose et al. (1995), Hawkins et al. (1996), and Brady et al. (1996).

The Kauffman permit stands apart from most surface mining permits in Pennsylvania by the abundance of geologic and hydrogeologic data that have been collected. During the application review period for this project, Hamilton and their consultants constructed 44 individual piezometers, drilled and analyzed the rock samples from 13 air rotary drill holes and six continuous core holes, and collected hundreds of samples from springs, ponds, streams, mine discharges, and others monitoring points. From this data researchers have arrived at a fairly good understanding of the local and intermediate flow systems that underlie the Kauffman permit (the regional flow system was not encountered). Nevertheless, there are still some anomalous data that have not been explained, so there is more to be learned from this site.

Geology

Stratigraphy - The geologic strata encountered during data collection for the Kauffman surface mine permit range from the middle of the Kittanning formation (hilltops) to the Mercer coal group (bottom of the deepest piezometers). Figure 2.24 is a generalized stratigraphic section of the Lower Kittanning coal to the Lower Mercer coal. The coal seam targeted for mining is the Lower Kittanning no. 3 seam, which serves as our reference horizon. The underlying Lower Kittanning no. 2 seam, separated from the Lower Kittanning no. 3 coal by a carbonaceous shale binder, is not of mineable quality (although it had been extracted over most of Phase I of the Kauffman permit). Above the Lower Kittanning no. 3 seam, the strata are commonly dark shales or siltstones which coarsen upward to the base of the Worthington sandstone. The Lower Kittanning no. 4 seam is not present on this site. However, at or near its stratigraphic horizon, approximately 10 feet (3 m) above the Lower Kittanning no. 3 seam, is a *Lingula*-bearing, dark gray to black shale of relatively high sulfur content. The Lower Kittanning no. 5 coal is referred to as the Lower Kittanning rider in the Kauffman permit file. Where present, it is a thin seam about 15 to 25 feet (4.5-7.5 m) above the Lower Kittanning no. 3 coal, near the base of the Worthington channel sandstone. The Worthington sandstone caps most of the high ground on the Kauffman permit. In places, this channel sandstone has encroached upon the Lower Kittanning no. 3 coal seam and replaced the coal in some areas. In the central and western portion of the Kauffman mine, the base of the Worthington sandstone may contain carbonate cements, calcite coatings on open fractures, and calcite fillings in "healed" fractures. The presence of this calcite has imparted some natural alkalinity to the deeper groundwater within the core of the hill.

Below the Lower Kittanning no. 2 coal are the Clarion no. 3, no. 2, and no. 1 coals in descending stratigraphic order. The Lower Kittanning no. 1 coal seam is sometimes present about 10 feet (3 m) beneath the Lower Kittanning no. 2, but that seam does not commonly occur on the Kauffman property. Between the Lower Kittanning no. 2 seam and the Clarion no. 3 seam, there lies a fairly persistent sandstone body of 5 feet (1.5 m) to more than 30 feet (9 m) in thickness. This stratum is designated the Kittanning sandstone and serves as a target aquifer for some of the piezometers. It appears that much of the strata among the Clarion coal seams are made up of shales, but much of the drilling below the Clarion no. 2 encountered so much water that identification of air rotary rock chips proved very difficult. Below the Clarion no. 1 seam and just above the Upper Mercer coal is the Homewood sandstone. The top of the Homewood sandstone served as the deepest aquifer encountered during piezometer construction.

Structure - Strata at the Kauffman site have an average dip of one to two degrees to the west-northwest. Stratigraphic and structural variations have imparted

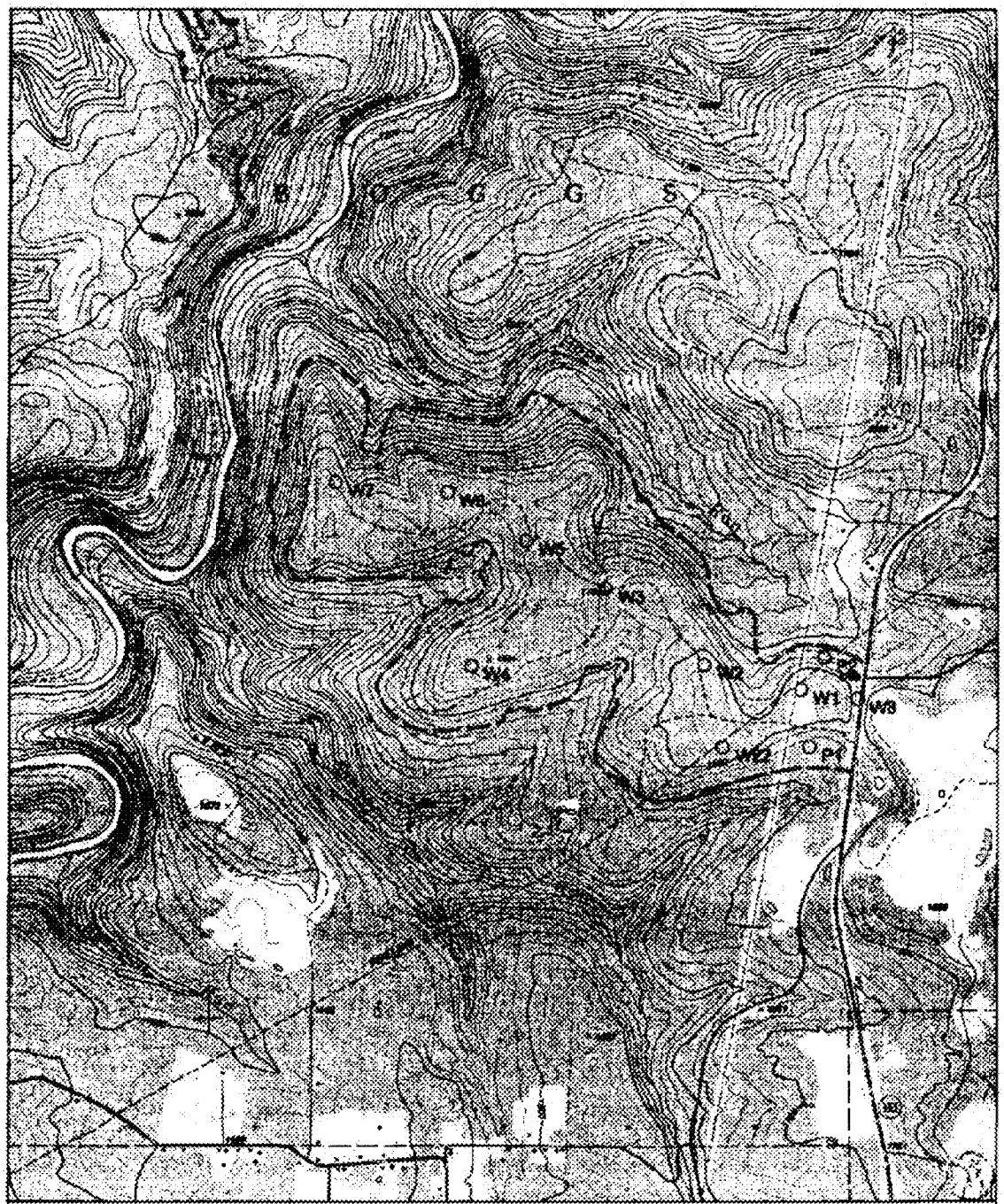


Figure 2.23 Kauffman site map.

some minor rolls to the coal surface, so dip directions and inclinations deviate slightly. In places the dip is very slight, less than one degree. In other areas, the dip exceeds eight degrees. According to Edmunds (1968), a major tear fault exists on the Thompson permit area just east of State Route 0153. However, its existence is only inferred on the Kauffman site. This fault strikes northwest-southeast and accounts for a stratigraphic offset of more than 295 feet (90 m) near the headwaters of Coal Run (about four and one half kilometers southeast of the Kauffman permit). In the eastern portion of the Kauffman permit, Edmunds (1968) infers that the fault could offset the strata by about 40 feet (12 m). However, pit-floor surveys on

Phase I of the Kauffman permit did not reveal displacement of the Lower Kittanning coal seam. Bedding-plane faults, however, do occur on the Kauffman site. These are evidenced by deformed bedding. The faulting east of the site may have dissipated at the Kauffman site as bedding-plane faults.

Jointing - Nickelsen and Williams (1955) indicate that the systematic joints in rock above the coal seams strike between N30°W and N45°W in the Houtzdale-Philipsburg area and that nonsystematic joints strike between N45°E and N60°E. These directions correspond to the general trend of drainage in the area of the Kauffman mine. Site specific studies at the Kauffman permit have not been done to confirm these joint orientations.

Hawkins et al. (1996) refers to the importance of stress-relief fractures. His discussion of Appalachian Plateau shallow groundwater flow is based to a large extent on observations at the Kauffman site. Bedding-plane or horizontal fractures are supposedly more common to the valley bottoms where rocks are under horizontal compressional stresses. However, Hawkins observed that horizontal fractures on the uplands of the Kauffman site were very common as well. Unlike tectonic joints, stress-relief fractures are associated with the surface and decrease in frequency with depth. Beyond the realm of the stress-relief fractures, the fracture frequency drops off considerably and groundwater movement is much more impeded.

Hydrogeology

Groundwater monitoring - Over 40 piezometers were constructed on the proposed Kauffman mine site as part of the mining company's application for a surface mining permit. Additionally, the operator drilled open boreholes to determine hydrogeologic properties of individual strata overlying the coal to be mined. Many piezometers were developed in clusters in which the individual wells were drilled about 6.5 to 13 feet (2 to 4 m) apart. There were nine main piezometer clusters of three, four, and five piezometers each. These piezometer clusters were identified as W1 through W8, and W22. Other piezometer locations included P1 and P2 for examining the pollutive groundwater plume on the flanks of the hill in the eastern Kauffman permit area, and SW1, SW2, and SW3 for monitoring shallow groundwater at certain permit locations.

Piezometers were constructed using four-inch (11.4-cm), thread-flush, Schedule 40 PVC pipe. Only the "AS" wells (this letter identification of piezometers is explained in the following paragraph) were constructed from one-inch (2.8-cm) pipe. All piezometers were installed with a 5.5-foot (1.7-m) section of slotted pipe (screen) at the base, except for the "C" wells which had an 11-foot (3.4-m) section of screen. The construction method was to backfill the annular region around the screen with sand to the top of the screen. The height of the sand pack for the "E" wells varied and sometimes extended above the screen by as much as 10.5 feet (3.3 m). Above the sand pack, rounded bentonite pellets were poured into the annular spacing to a pre-swell height of about 5.5 feet (1.7 m). The depth to the top of these construction layers was constantly checked with a weight and steel tape. The rounded bentonite pellets were used because they were less likely to bridge between the side of the borehole and piezometer pipe when the hole was wet or flooded. After the rounded pellets were poured in, the rest of the hole was completely filled with bentonite chips (Volplug®). No drill cuttings or dirt was pushed in. This further reduced the chance of an artificial connection developing between aquifers via the annular space. A steel cover, set in concrete, was then placed at the surface to protect the plastic casing. This cover is kept locked until the sampler needs to access the piezometer.

Each piezometer within a well cluster was identified by a letter indicating the target horizon being tested. Figure 2.24 shows the generalized stratigraphic section and the corresponding piezometer target zones. Piezometers designated as "A" wells were constructed so as to monitor the Lower Kittanning no. 2 and no. 3 coal seam aquifer. At this sampling horizon, a 5-foot (1.5-m) screen was installed to monitor the groundwater pressures. Inside the boreholes used to construct the A wells, the operator installed an "AS" level piezometer of smaller diameter down to the base of the Worthington sandstone. The level of the AS screen with respect to the "A" piezometer screen varied as the depth of the Worthington sandstone varied. Next, the "B" wells were installed and were usually open (5-foot (1.5-m) screen) to the Clarion no. 2 coal horizon. One exception was piezometer W1B, which was finished to the shallower Clarion no. 3 horizon. The "C" wells were open (10-foot (3-m) screen) between 9 and 15 meters below the B level screens, either at the Clarion no. 1 coal horizon or the top of the Homewood sandstone just beneath that seam. Generally speaking, piezometers W1C, W2C, W3C, and W4C were screened at about the Clarion no. 1 coal horizon, and the C-level piezometers at well pods W5C, W6C, W7C, W8C,

and W22C were screened just below the Clarion no. 1 within the Homewood sandstone. In most locations, the drilling became so wet below the Clarion no. 2 horizon that logging of the strata was very difficult. Evidently, a significant aquifer(s) exists beneath the Clarion no. 2 underclay. Finally, the wells designated as "D" wells were intended as open boreholes, not true piezometers, and were developed to conduct packer pumping tests on various strata at and above the main coal.

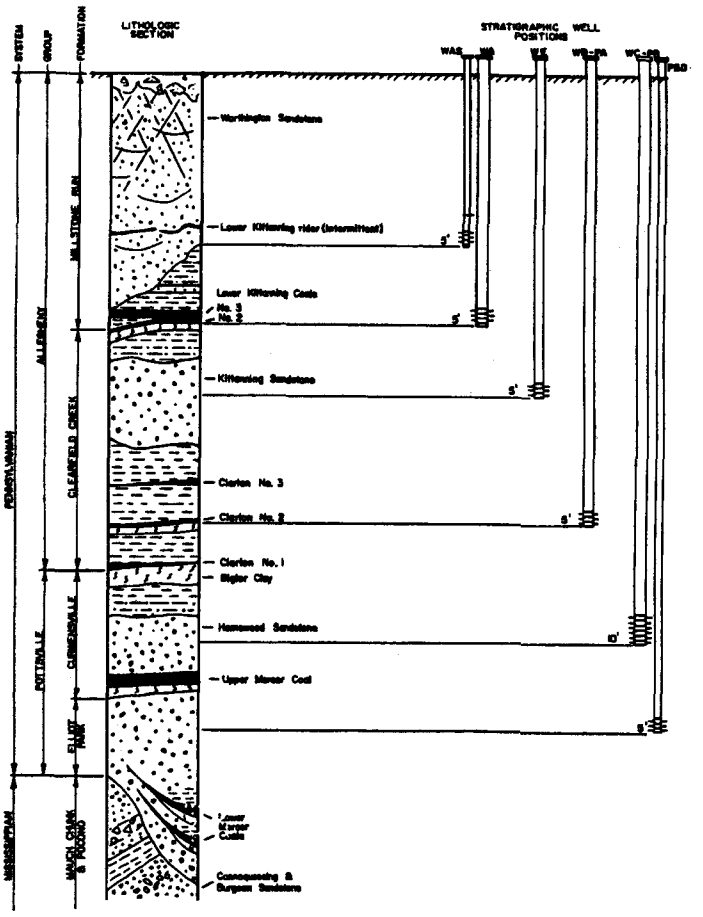


Figure 2.24 Generalized geologic section with piezometer cluster schematic.

After the initial piezometers were installed, a fifth piezometer was drilled at well pods W1, W2, W6, W7, and W22. These "E" piezometers were completed into the Kittanning sandstone between the Lower Kittanning no. 2 and Clarion no. 3 coal seams. These wells were developed to monitor the Thompson pollution plume. They monitored an unaffected groundwater horizon between the Lower Kittanning coal that was to be mined under the Kauffman permit and the underlying pollution plume. The "E" piezometers monitored for possible downward migration of AMD through the new Lower Kittanning pit floor. Other piezometers were installed to monitor the Thompson pollution plume on the flanks of the ridge within the eastern portion of the Kauffman site. They were designated as "PA", "PB", and "PBD" wells, and were screened at the Clarion no. 2, the Homewood sandstone, and the Upper/Lower Mercer coal interburden, respectively.

Groundwater flow within the weathered zone - Groundwater flow on the Kauffman site is controlled by numerous natural factors including topography, relief, stratigraphy, dip, joint sets, bedding-plane separations, zones of fracture concentration, and stress-relief fractures. Additionally, man-induced influences such as road excavations, surface mining, underground mining, and numerous exploratory drill holes have also influenced the groundwater flow.

Hawkins et al. (1996) described the relatively shallow groundwater system at the Kauffman site as fracture flow. Two distinct hydrogeologic settings through which most groundwaters flow were described: 1) a highly fractured weathered zone near the surface, and 2) a less fractured, non-weathered zone at depth. The weathered zone is characterized by numerous stress-relief fractures and bedding-plane separations that become less abundant with increasing depth. This abundance of stress-relief fractures provides a highly transmissive environment for groundwater flow. Underlying the weathered zone, the rock is less densely fractured due largely to the paucity of stress-relief fractures. Data and observations by Hawkins et al. (1996) do not show a sharp transition from the weathered zone to the unweathered zone. Rather, the frequency of stress-relief fractures diminishes somewhat gradually as the depth increases. The "effective" depth of the densely fractured weathered zone varies with different topographic settings (hilltop, valley wall, and valley bottom) and from location to location within similar topographic settings. Hawkins et al. (1996) estimates a general depth for the weathered zone of about 65 feet (20 m) based upon his observations at Kauffman and actual measurements within highwalls of other mine sites. Kipp and Dinger (1987) noted that the depth of the weathered zone was usually 50 feet (15 m) on the hilltops and near 80 feet (24 m) along the valley walls.

Beneath the weathered zone, the main fractures conveying groundwater are the near-vertical systematic joints and concentrated fracture zones developed from tectonic stresses. Unlike stress-relief fractures, these tectonic fractures are not depth related. They can extend from the surface to depths of several hundred feet.

However, at considerable depth their ability to transmit groundwater is usually much reduced. Due to the conductivity contrast, the unweathered rock underlying the surface and near-surface, weathered zone behaves as an aquitard and maintains a perched water table within the weathered fracture zone. Groundwater movement within this near-surface "aquifer" generally follows the topography and migrates toward the surrounding hillslopes and adjacent valley floor. Where a confining layer such as a coal underclay is present, groundwater may be diverted to the surface as contact springs along the hillside.

Not all groundwater within the stress-relief fracture zone is directed laterally toward the stream valleys. The aquitard zone, which serves to perch water, only represents a relative contrast in hydraulic conductivity and is not an impermeable barrier. Therefore, a significant amount of groundwater will leak vertically into the underlying strata. Vertical fracture zones of tectonic origin enhance this downward leakage in areas adjacent to such features. Also, abandoned boreholes that have remained open will present a means of quickly conveying shallow groundwater into the deeper systems.

Groundwater that emanates from a particular stratigraphic horizon along the hillside is not necessarily associated with groundwater at that same horizon within the core of the hill. Water chemistry data of contact springs sampled along the unmined hillslopes of the Kauffman property were compared to water chemistry data collected from deeper hilltop piezometers. The more dilute chemistry of the hillside springs is thought to represent groundwater of the weathered-rock zone. In contrast, groundwater deep within the hill comes in contact with less weathered rock and moves much more slowly, thereby increasing the contact time with the adjacent rock. Consequently, the deeper groundwater is higher in dissolved constituents.

At the Kauffman mine, a good example of this concept involves contact springs GR438 and GR439. Both apparently emanate from the Clarion formation just 175 meters (570 ft.) northwest and downdip of well pod W1. At well pod W1, piezometers W1B and W1C are open at the Clarion no. 3 and Clarion no. 1 coals. Both reveal the effects of AMD from the Thompson pollution plume. Initially it was thought this plume would spread north and west and reach springs GR438 and GR439. However, the springs do not exhibit characteristics of AMD. Apparently, the polluted water within the Clarion horizon at W1 "steps down" into deeper strata as the plume approaches the stream valley. The springs represent shallow groundwater from the weathered zone and are not necessarily representative of the "Clarion waters" within the hill core.

Groundwater flow within the unweathered zone - As discussed above, groundwater within the weathered zone flows laterally in accordance with the surface contours. This is due to the gradational contact between the weathered and non-weathered strata. This contact mirrors the topography. However, there also is a significant component of vertical leakage into deeper aquifer systems. As downward-flowing groundwater reaches stratigraphic layers of low vertical hydraulic conductivity relative to the adjacent strata, lateral flow occurs. Of course there will always be vertical leakage through each aquitard encountered, but the volume of vertical groundwater recharge is gradually reduced with increasing depth.

Initial monitoring of the Kauffman piezometers during the permit application review period confirmed that the general movement of groundwater is downward on the Kauffman site. The shallowest piezometers had the highest pressure head, the deepest piezometers had the lowest pressure head, and the intermediate piezometers fell somewhere in between. This is the textbook pattern for an upland setting that serves as a groundwater recharge area. However, certain well pods indicated that the groundwater flow patterns may be more complicated. Specifically, at well pods W3 and W6, the pressure heads for the deepest piezometers were equal to, or higher than, the pressure heads within the shallowest piezometers, maintaining very high water levels in the piezometer tubes. In fact, the water levels within the W3 and W6 piezometers were about 120 feet (37 m) and 111 feet (34 m), respectively, above the top of the screen.

Hydrographs for the W3 and W6 piezometer pods are provided as Figures 2.25 and 2.26, respectively. The hydrograph for the W3 pod includes several years of weekly water level measurements at piezometers W3AS, W3A, W3B, and W3C. For the first couple of years, the water levels within the W3C piezometer were elevated above the level of the W3A piezometer. During the permit application review period, the permittee's consultants interpreted this as localized confining conditions. However, subsequent data revealed that pressures fluctuate greatly at the W3C horizon, and for much of the time since 1994, the groundwater flow pattern has been more typical of the "normal" well pods. Also, the water level plots for the W3AS,

W3A, and W3B piezometers exhibit similar seasonal and climatic patterns. The mimicry of the lows and highs suggests that these aquifers are hydrologically connected. On the other hand, there appears to be no such relationship between W3C and the other three piezometers.

Thus far, there has been no complete aquifer testing of the W3 piezometers. We do not yet know how each screened zone responds to pumping and/or slug testing, or even how rapidly the wells recharge after being pumped dry. Nonetheless, recharge data from the well sampler's records provide indications of the relative transmissivity among the water-bearing zones adjacent to each piezometer screen. For instance at W3C, the deepest piezometer of the W3 pod, recharge is very slow after pumping - probably less than 0.05 liters per minute (L/m) (0.013 gpm). On the other hand, recharge is relatively rapid, probably more than 3.5 L/m (0.92 gpm), within the intermediate-level W3B piezometer. The shallow W3A piezometer recharges at a relatively moderate rate of about 1.5 L/m (0.39 gpm).

The hydrograph for the W6 well pod reveals a much different response in the deepest piezometer, W6C, as compared to W3C. Soon after W6C was developed, the water level climbed to just below that of W6A, and it has mimicked very closely the hydrograph of that piezometer ever since. In fact, piezometers W6A, W6C, and W6E are all closely associated with respect to water elevation and hydrographs. Furthermore, the W6B hydrograph also mimics the other three piezometers, but does not maintain such a high water level. This elevation difference between the water levels within the W6B piezometer and the other three piezometers was initially over 46 feet (14 m), but it has been reduced recently to about 33 feet (10 m).

Interestingly, as noted on the W6 hydrograph, the water levels in all four piezometers had dropped following the excavation of an open cut on the Lower Kittanning coal seam about 900 feet (275 m) to the west-northwest of the well pod. The floor of this mining cut, in addition to being relatively far from the W6 piezometers, was about 100 feet (30 m) higher in elevation than the stratigraphic horizon open to the W6C piezometer screen, and about 62 feet (19 m) above the horizon open to the W6B screen. Nevertheless, these deeper aquifers still responded to the excavation of the open mining pit.

The mimicry among the four aquifers tested at the W6 well pod (and among three of the four piezometers at the W3 well pod) suggests there is a vertical hydrologic connection, either natural or artificial. As for an artificial connection, the most likely explanation would be poor piezometer construction. For example, if the piezometer was not backfilled properly, groundwater from the upper aquifers may drain to the screen through the annulus. Another possibility could be a defect in the piezometer casing, such as a crack or separation at the joints, allowing shallower groundwater to seep into the piezometer above the screen. In either case, bentonite seals would have been compromised. This seems unlikely since bentonite pellets and chips were used from the top of the screen to the surface. Furthermore, because more than one piezometer exhibits a mimicry, each piezometer would have to be faulty to explain an artificial connection in this manner.

Artificial connection of aquifers can also occur through open and abandoned exploratory boreholes. Although the Kauffman property is riddled with exploratory boreholes down to the Lower Kittanning coal seam, there are probably few, if any, old boreholes as deep as the Homewood sandstone. Therefore, it seems to be more likely that the connection among the various aquifers is natural rather than artificial.

As for a natural connection, the aquifers could be connected through tectonic fractures, zones of fracture concentration, or fault zones. If discrete fractures or fracture zones are present at or very near well pod W6, vertical communication between the aquifers at that location could be established. Such fractures or fracture zones could serve as avenues for groundwater recharge to deeper aquifers.

As with the W3 well pod, the sampler's records during pumping and sampling at the W6 pod reveal some qualitative aquifer characteristics. The deep W6C piezometer recharged the quickest, approximately 2.8 L/m (0.74 gpm). In comparison, intermediate piezometer W6B recharges the slowest, approximately 0.4 L/m (0.1 gpm) while piezometer W6E recharged at a moderate rate of approximately 0.9 L/m (0.24 gpm). Although the sampler's records provide some qualitative information regarding the relative transmissivities among the aquifers, quantitative pumping tests, slug tests, and recovery tests should be conducted in the future if further study is desired.

Although the majority of Kauffman piezometers reveal a fairly straightforward groundwater regime, well pods W3 and W6 reveal certain aberrations that may

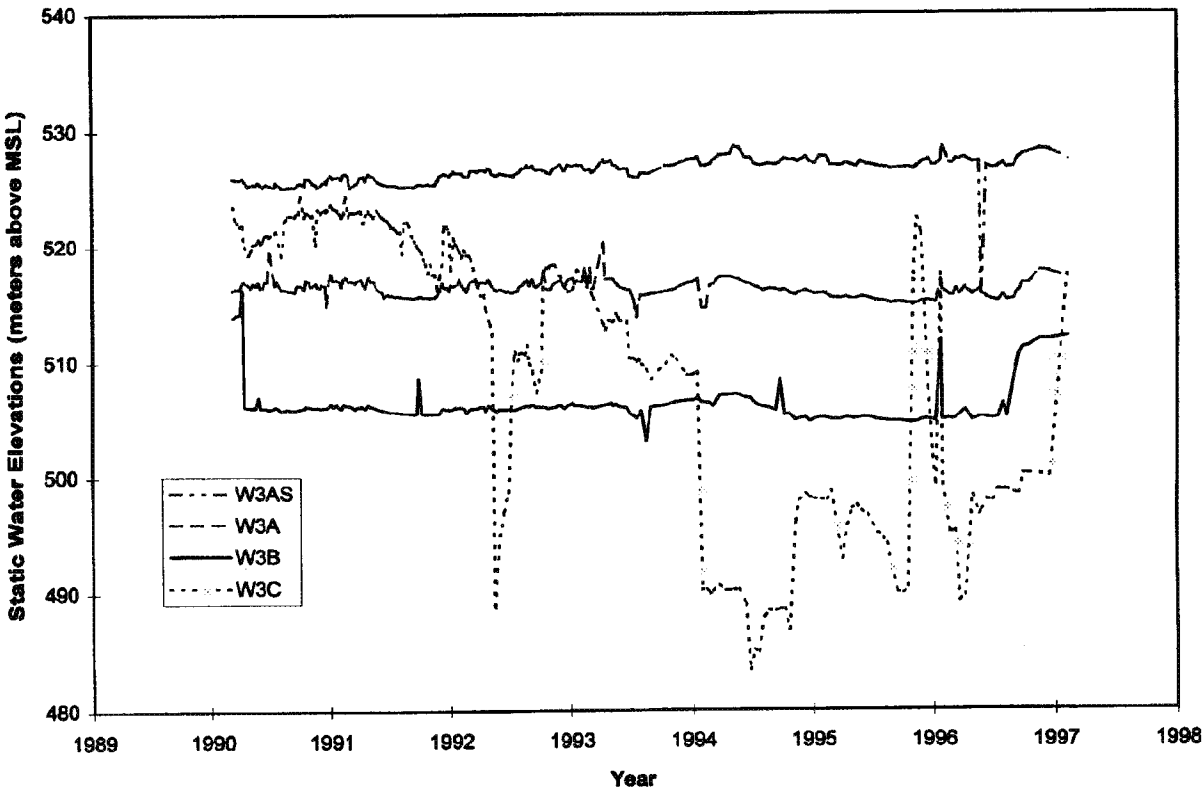


Figure 2.25 Hydrograph of W3 piezometer pod representing weekly water level measurements from March 1990 through February 1997.

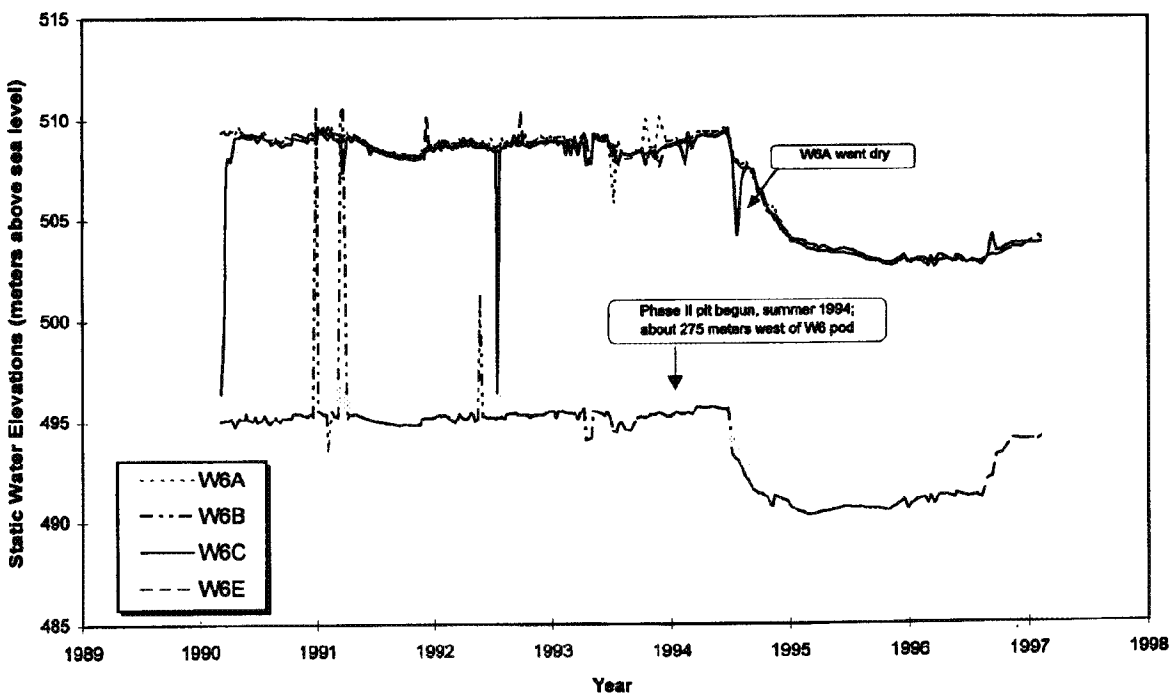


Figure 2.26 Hydrograph of W6 piezometer pod representing weekly water level measurements from March 1990 through February 1997.

be common where significant fracturing of the rock mass is a factor. Such fractures may provide the vertical avenues for interconnecting perched aquifers that would otherwise be hydrologically isolated. The mimicry in hydrograph traces among separate piezometers at various levels can hardly be ignored. The piezometers at well pod W6 present an especially striking example.

Summary - Although tectonic fracturing complicates groundwater flow patterns, the general movement of groundwater on the Kauffman site is fairly typical of an upland setting within the Allegheny Plateau. Stress-relief fracturing near the surface establishes a relatively shallow water table aquifer that responds quickly to precipitation events and directs groundwater flow laterally toward the adjacent stream valleys. Groundwater within this weathered zone has short residence time from infiltration to discharge and therefore contains less concentrations of dissolved minerals than does groundwater deep within the core of the hill. Along the hillsides, this shallow groundwater generally does not emerge from beneath the colluvium but provides baseflow to wetlands and streams on the valley floor. Nevertheless, a confining layer such as a thick clay stratum may divert flow to the surface as a contact spring or seepage face along the clay seam outcrop. Such contact springs that originate from the weathered zone are not representative of groundwater at the same horizon within the core of the hill.

Although groundwater within the weathered zone generally flows in accordance with topography, there is significant vertical leakage to the underlying strata, primarily through joints. This vertical leakage is further enhanced by the presence of faults and major fracture zones. Where these percolating waters reach strata of contrasting vertical conductivities, additional flow zones are established at depth. Flow within these deeper aquifers is again directed laterally but is more controlled by structural features such as dip, joints, fracture zones, and faults. Of course, stratigraphic variations in the aquifer or underlying aquitard will also influence the flow. Each of these aquifers that are established at depth will leak to the underlying strata, but the volume of such leakage will gradually diminish as well the occurrence of joints through which the groundwater flows.

Most piezometers at the Kauffman site confirm this downward movement of groundwater that is typical of a recharge area. However, a couple of piezometer clusters or pods suggest a vertical connection among the various aquifers that are monitored. These piezometers do not show the normal pattern of decreasing pressure head with each lower aquifer, and some reveal a close mimicry among hydrograph plots. Assuming that the piezometers are constructed properly, this interconnection may be caused by the presence of a significant tectonic fracture zone or a fault, but confined groundwater conditions cannot be ruled out either.

Additional study is needed on the Kauffman site to further examine the apparent anomalies in groundwater flow patterns. Thus far, no pumping tests or slug testing has been conducted on the two piezometer pods where unusual water levels were recorded. Further research may provide some additional insight into the behavior of groundwater flow within this type of setting on the Appalachian Plateau.

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