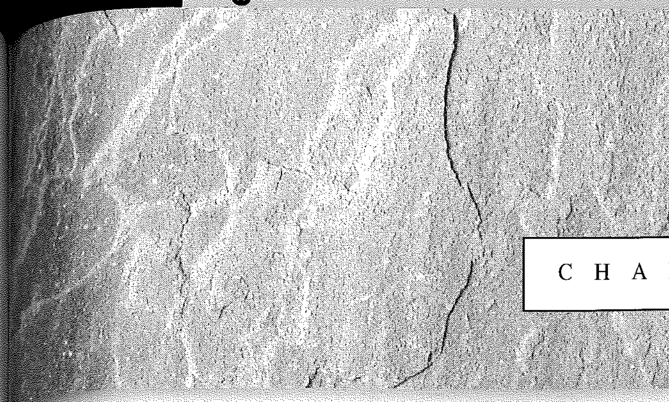


CHAPTER 8

ABNORMAL FLUID PRESSURES

8.1. OVERPRESSURES AND UNDERPRESSURES.

Abnormal fluid pressures are those that are above or below hydrostatic. **Hydrostatic** fluid pressures are those in which the fluid pressure at any depth is due to the weight of the overlying fluid (as defined by equation 2.23). Nominal hydrostatic fluid-pressure gradients are usually about 10.5 kPa-m^{-1} (0.465 psi-ft^{-1}). **Lithostatic** pressure is the pressure due to the weight of the entire overburden (fluid plus matrix). Fluid pressures generally cannot exceed lithostatic, as fluid pressures in excess of lithostatic cannot be contained by the total overburden weight. Because rocks have lateral strength, however, it is possible to find isolated occurrences of fluid pressures which are slightly in excess of lithostatic.

Fluid pressures below hydrostatic are termed **underpressures.** Fluid pressures in excess of hydrostatic are termed **overpressures** or **geopressures**. Overpressures in sedimentary basins tend to be more common than underpressures. Sedimentary basins with overpressures typically consist of hydrostatically pressured sediments extending from the surface to depths of 2 to 3 kilometers. The hydrostatic section is underlain by a transition interval, followed by a deep section of abnormally high fluid pressure and fluid-pressure gradients (Figure 8.1). There is considerable interest in understanding the origin and evolution of overpressures, as abnormally high pressures represent a hazard in drilling for petroleum. Under normal circumstances, high fluid pressures at depth are balanced by the weight of the drilling fluid in the wellbore. Drilling engineers usually tend to make the overall weight of the drilling fluid column slightly higher than is necessary so as to avoid a catastrophic blowout. If a zone of abnormally high fluid pressure is unexpectedly encountered, there is danger of a destructive blowout wherein formation fluids rush up the wellbore at great speeds.

8.2. STATIC VERSUS DYNAMIC HYPOTHESES.

As discussed by Bredehoeft et al (1994), there are two distinct schools of thought on the creation and maintenance of anomalous fluid pressures in the Earth. The **static school** (Bradley, 1975; Hunt, 1990; Bradley and Powley, 1994) believes that abnormal pressures, regardless of origin, are maintained by pressure seals (Hunt, 1990; Ortoleva,

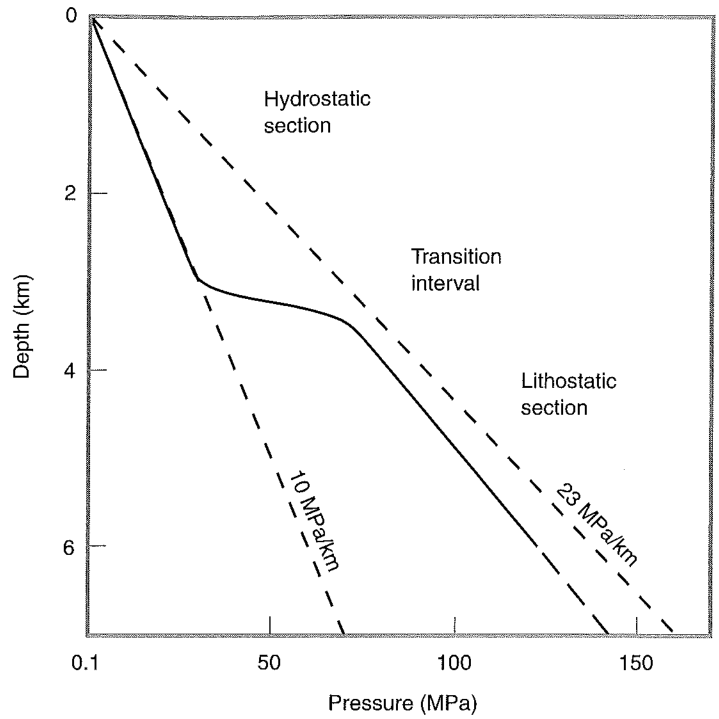


Figure 8.1 General trend of fluid pressure versus depth in basins with Gulf Coast-type geopressures. (From Bethke, 1986, p. 6536.)

1994a, b) (Figure 8.2). A **pressure seal** was defined by Hunt (1990, p. 2) as (emphasis added):

> . . . a zone of rocks capable of hydraulic sealing, that is, preventing the flow of oil, gas, and water. The term does *not* refer to capillary seals . . . the term refers to seals that prevent essentially *all* pore fluid movement over substantial intervals of geologic time.

Other authors have implicitly defined the term "pressure seal" more loosely by using it to refer to any low-permeability unit, however, as the above quote from Hunt (1990) shows, the original intention was to apply the term to units that essentially behave as if they have zero permeability over substantial intervals of geologic time (10^7-10^8 yrs).

Pressure seals are one aspect of a paradigm wherein it is thought that sedimentary basins have two superimposed hydrogeological systems: a shallow system characterized by hydrostatic pressures, and a deeper system consisting of a series of overpressured, hydraulically isolated pressure compartments (Hunt, 1990; Bradley and Powley, 1994) (Figures 8.3 and 8.4). A **pressure compartment** is a three-dimensional hydraulically isolated volume of the Earth's crust that has a fluid pressure different from the ambient surroundings. The role of the pressure seal is to maintain anomalous (above hydrostatic) pressures in the lower system by preventing the movement of fluid across compartment boundaries. Compartments may be breached by fracturing when fluid pressures exceed lithostatic, but the seal regains its integrity when the fluid pressure drops below lithostatic.

There are several difficulties with the pressure-seal pressure-compartment concept. One is the lack of a known geochemical/geologic mechanism to create pressure seals. Some authors have claimed that pressure seals exhibit distinct diage-

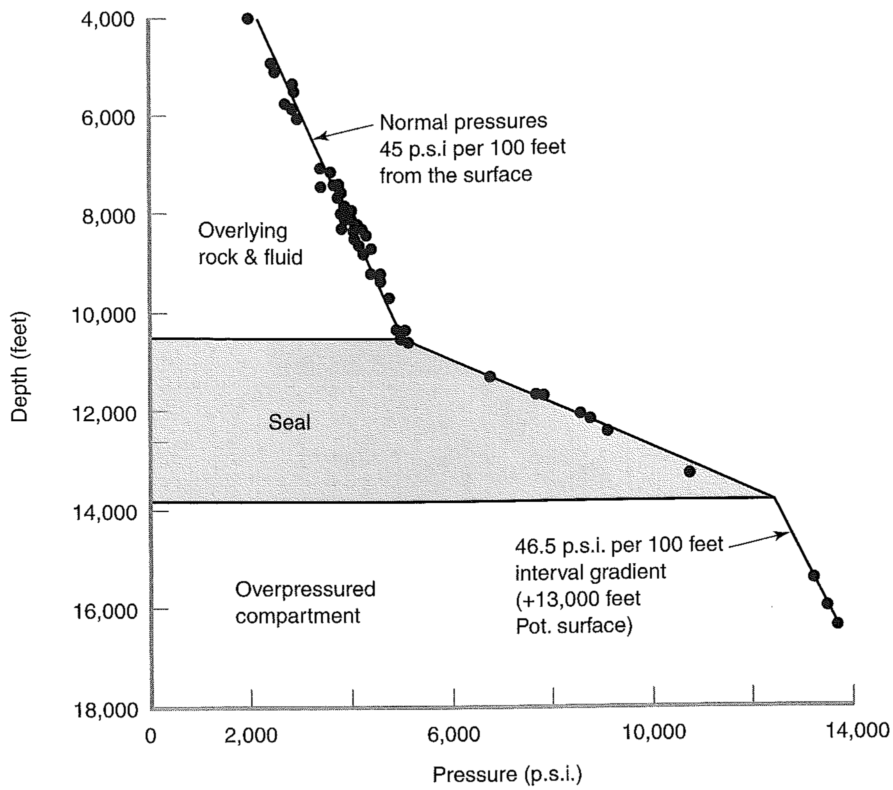


Figure 8.2 Fluid pressure as a function of depth, Cook Inlet fields, Alaska. Pressure seal is inferred to exist in region of high fluid pressure gradient.
(From Hunt, 1990, p. 6; after Powley, 1980.)

netic bands (Al-Shaieb et al., 1994), suggesting the possibility that diagenetic processes such as calcite precipitation in shale pore spaces may be an effective sealing mechanism. But, this idea remains largely untested. As Bradley and Powley (1994) noted, "*no* example of a directly measured, permeability-defined seal is known." The existence of pressure seals is inferred from differences in hydraulic potential measured across relatively permeable reservoir rocks. Even if top- and bottom-bounding pressure seals could come into existence through diagenetic processes, it is difficult to imagine what geologic features could function as lateral seals. Faults have been invoked as likely candidates, but faults tend to be conveniently invoked alternately as either seals or conduits for flow as suits the need of specific circumstances.

Unlike Hunt (1990) who explicitly specified that pressure seals are *not* capillary seals, Revil et al (1998) proposed that sealing could occur in sedimentary basins through the formation of capillary gas seals. A capillary force is an attractive force that exists between two different substances (see section 6.1.1). Rock pores and channels that exist between solid matrix grains will act as capillary tubes, drawing water into them and holding it there until forcefully displaced. Thus a fine-grained sedimentary rock saturated with water may act as a layer of zero permeability to prevent the entry of fluids such as oil or gas unless the capillary force can be overcome. If alternating layers of fine- and coarse-grained sediments are present, gas may accumulate in the coarse-grained sediments with capillary forces preventing it from entering into adjacent

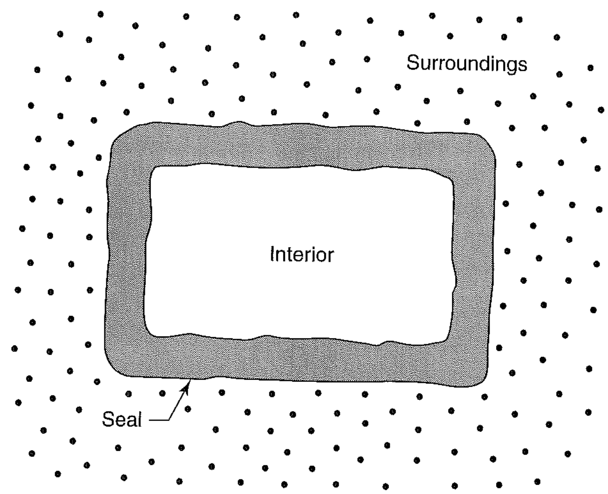


Figure 8.3 Conceptual model of generic pressure compartment. Interior of compartment is hydrostatically pressured, and is of relatively high-permeability. The interior is separated from its surroundings by a pressure seal.
(From Ortoleva, 1994b, p. 40.)

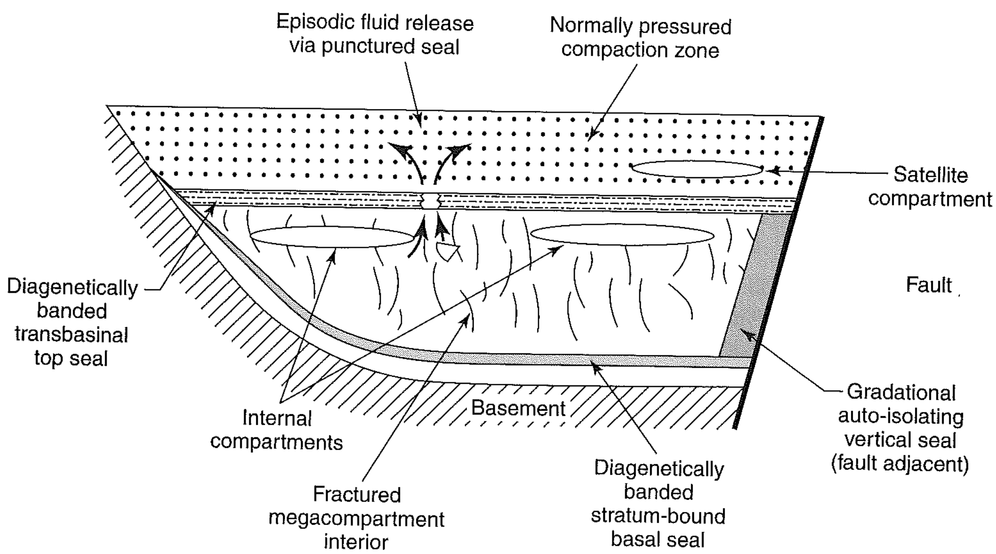


Figure 8.4 Pressure compartment/seal paradigm of sedimentary basin hydrogeology showing different levels of compartmentalization.
(From Ortoleva, 1994b, p. 44.)

water-saturated fine-grained rocks. The immobile gas thus constitutes a seal of zero permeability—unless the hydraulic gradient is large enough to overcome the capillary force.

A hypothesis that invokes sealing by gas capillary forces has several advantages. As gas generation also tends to produce abnormally high fluid pressures, the hypothesis is parsimonious in that it simultaneously provides a mechanism for both generating and maintaining anomalous fluid pressures. Capillary sealing also provides a mechanism for achieving zero permeabilities without

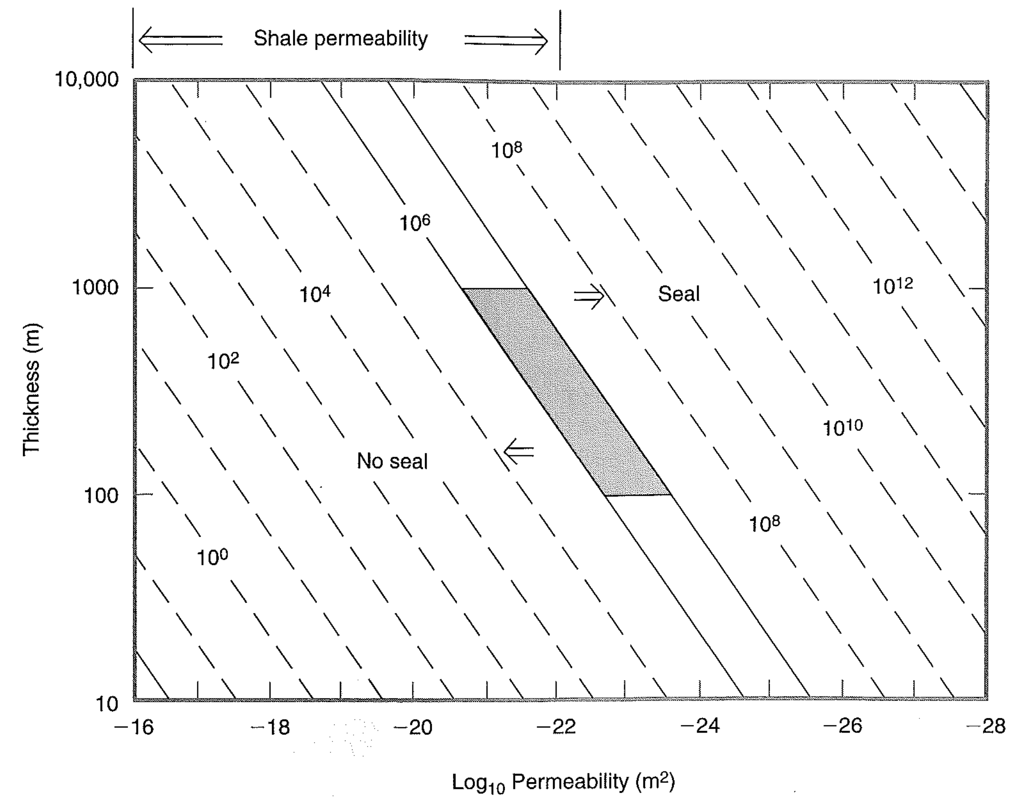


Figure 8.5 Maximum time (in years) over which a layer of given thickness and permeability may confine excess pressures. Shaded area indicates approximate permeability required to sustain a 100 to 1000-m-thick seal over geologic time (about 1 Ma).
(From Deming, 1994a, p. 1008.)

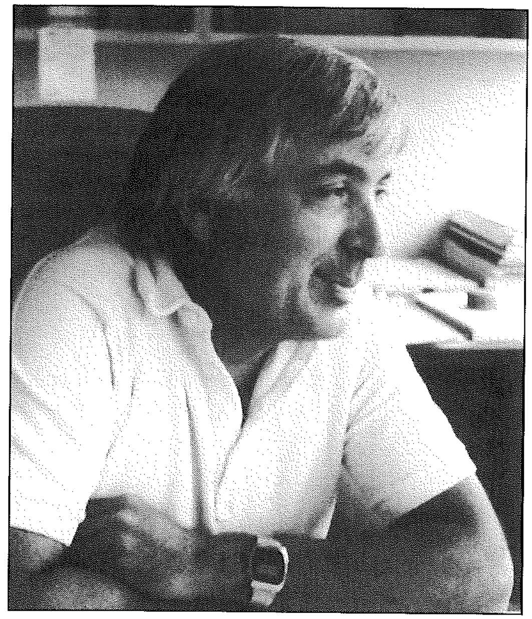
contradicting established dynamic paradigms, which maintain that aquicludes are rare to nonexistent. For example, Neuzil (1994) reviewed the permeability of shales and clays and found that most argillaceous media have permeabilities greater than 10^{-20} m^2, too high by three to four orders of magnitude to preserve abnormal fluid pressures for 100 Ma. Capillary sealing also explains how pressure compartments can be sealed in all three dimensions. Gas follows fluids along the path of least resistance until all escape routes are plugged. Finally, sealing by capillary forces can produce the type of compartmentalization apparently observed in basins such as the Anadarko Basin in southwest Oklahoma. The existence of levels of pressure compartmentalization is difficult to explain unless some type of physical or chemical sealing-mechanism is invoked.

Deming (1994a) quantified the conditions under which pressure seals may retain abnormal pressures by calculating the characteristic time a seal of a specified thickness and permeability may confine a pressure transient (review section 5.5). He found that to confine abnormal pressures for more than 10^8 yrs with a seal 100 to 1000 m thick would require seal permeabilities in the range of 10^{-23} to 10^{-25} m^2 (Figure 8.5). This range is near or below the lowest extreme of measured shale permeabilities (Figure 3.14). It is thus difficult to maintain abnormal fluid pressures over geologic time without either pressure seals or the influence of an active and ongoing geologic process to offset

John D. Bredehoeft: Fluids in Geologic Processes

by L. F. Konikow and P. A. Hsieh

U.S. Geological Survey, Menlo Park, California



John D. Bredehoeft

John D. Bredehoeft (born in 1933) is a quantitative hydrogeologist who has consistently pushed outward the frontiers of the science. He is still active and has a long productive history of accomplishments, which have won him many awards and international recognition. He has a talent for identifying critical problems, simplifying each down to a tractable question, and then deriving and publishing a solution that has great implications and transfer value. Some of his work has been earthshaking. Literally!

John received his undergraduate education at Princeton University, where his major was geological engineering. His graduate studies were at the University of Illinois, where he received an M.S. and Ph.D. in Geology. Reflecting on his early career, John remarked (after receiving the Horton Medal from the American Geophysical Union in 1997),

> I was lucky to go to the University of Illinois, where my major professor and mentor was Burke Maxey. He instilled in those of us who were associated with him a demand for excellence. Upon receiving my Ph.D. in the early 1960s, I was lucky to go to work at the U.S. Geological Survey. I arrived at a time when I could apprentice with some of the best professionals engaged in the study of groundwater.

John's remarks clearly reflect the generosity of his outlook on doing scientific work. He shares credit and he shares ideas. And those who have been lucky enough to be mentored by John know that his ideas are usually gems. Many recipients of his generous support and encouragement of young scientists have carried on the tradition of excellence of work and generous sharing of ideas.

John's work often reflects a multidisciplinary approach to solving difficult problems. He has made important advances linking groundwater hydrology with geophysics, geochemistry, tectonics, petroleum engineering, economics, and numerical methods. John's interests and work have great breadth and depth, as they extend beyond the purely technical realm of science and engineering into the social realms of management of natural resources, the management and administration of research organizations, and even philosophy of science.

A common thread running through much of John's work is his interest in the role of fluids in geologic processes. John's first publication, in 1963, was the first quantitative examination of membrane filtration in the subsurface. His 1968 papers on anomalous fluid pressure were the first cogent analyses of geologic processes as hydrologic driving forces and the first recognition of anomalous pressures as hydrodynamic transients. His 1967 paper on the response of aquifers to earth tides is extensively cited as the seminal paper on that topic. His analysis of thermal profiles for estimating groundwater flow rates is elegantly simple, yet has proven to be of tremendous utility. He was among the first to use hydraulic fracturing as a method for determining the state of stress in the subsurface. In 1970, many geologists barely recognized the existence, let alone the importance, of subsurface fluids. That is no longer true today, and geologists in great numbers are now looking at how groundwater controls or influences ore deposits, hydrocarbon reservoirs, tectonic processes, volcanic events, and almost every other sub-field of geological and geophysical sciences.

During the time that John was creating and applying new computer simulation models of groundwater processes, he was also "shaking things up" as a participant in the well-known Rangely, Colorado, experiments (where earthquakes were created and controlled by high-pressure fluid injection). He followed up on this by contributing to the Parkfield, California, earthquake studies where he was a proponent of using water wells as strain meters to monitor deformation of the earth near faults, partly in search of potential earthquake precursors.

In the realm of groundwater systems analysis, John has made several fundamental contributions to methods of well test analysis. He was instrumental in the development of the rigorous theory of slug tests, now one of the basic tools of the field hydrogeologist. He also extended the slug test technique to solve the difficult problem of field measurement of very low permeabilities.

Most practicing hydrogeologists today routinely apply computer simulation models to help them understand and solve the particular problem being addressed. They all owe a debt of gratitude to John Bredehoeft, who helped pioneer the development and application of digital simulation of groundwater systems when most hydrologists were still using analog models. His papers, particularly those co-authored with George Pinder, were widely recognized as standard references in groundwater model analysis. Many model developers built upon the basics that John laid out, and many of today's flow and transport modelers use programs based on his work. In the early 1970s, John was among the few who saw the significance and pervasiveness of groundwater contamination problems: this was a motivating factor for his pushing strongly for the development and application of solute-transport models. Less than 20 years later, dealing with groundwater contamination had become a multi-billion dollar a year industry.

John wanted society to benefit from tax-supported research. His interest in promoting a "practical payoff" of science is illustrated in the area of groundwater management, where he showed how economic theories can be applied in light of realistically variable hydrogeologic conditions to develop policies for water allocation or development of groundwater resources. John analyzed topics such as groundwater depletion, conjunctive use, and artificial recharge. In countering what he considered a common "groundwater myth," he demonstrated the fallacy of basing groundwater management rules (such as restrictions on pumpage) on computed water budgets (or recharge rates) for conditions prevailing prior to development.

John is not only a leading scientist, but a leader of scientists. John served for a number of years as Chief of the National Research Program of the Water Resources Division (WRD) of the USGS, which at that time employed close to 300 scientists and engineers. In this position, John substantially increased the relevance and visibility of this hydrologic research program. He later served for several years as the Regional Hydrologist for the operational program in the eight-state Western Region of WRD. What is perhaps the most amazing feat is that John remained a productive scientist and researcher during the years he served as a manager.

For Additional Reading

Freeze, R. A., Bredehoeft, J. D., and Pinder, G. F. 1976. Presentation of the O. E. Meinzer Award to John D. Bredehoeft and George F. Pinder, Citation and Responses. *GSA Bulletin,* 87 (8): 1212–1213.

Konikow, L. F., and Bredehoeft, J. D. 1998. Bredehoeft Receives Robert E. Horton Medal, Citation and Response. *EOS, Transactions, American Geophysical Union,* 79 (8): 101.

Wolff, R. G., and Bredehoeft, J. D. 1997. Penrose Medal presented to John D. Bredehoeft, Citation and Response. *GSA Today,* 8 (3): 13–14.

the natural tendency for pressure equalization. If pressure seals exist, it thus seems likely that some type of active geochemical or capillary mechanism is necessary. Otherwise, rock permeabilities are simply too high to allow high fluid pressures to exist in the Earth's crust over geologic time.

The **dynamic school** embraces the classical hydrogeologic tenet that "there are no totally impermeable geologic materials" (Bredehoeft et al., 1982, p. 297) (with, of course, rare exceptions for materials such as permafrost and salt). Pressure seals simply do not exist in this paradigm. Abnormal formation pressures may be caused by a transient disturbance related to some ongoing geologic process (e.g., rapid sedimentation), or an equilibrium process such as topographically driven fluid flow (Tóth, 962; Neuzil, 1995). The existence of abnormal pressures is seen as an indication that the rate of pressure generation (either positive or negative) is sufficiently high so as to maintain abnormal pressures in the presence of low-permeability rocks for substantial periods of geologic time.

8.3. Causes of Abnormal Fluid Pressures.

Hypothetical mechanisms for the creation of abnormal fluid pressures in the Earth's crust can be either steady-state or transient. Generally speaking, there is only one steady-state mechanism, but inumerable transient processes that can lead to abnormal fluid pressures.

8.3.1. A Steady-state Mechanism.

Topographically driven flow is a steady-state mechanism that can lead to both under- and overpressures. As long as the topography and permeability remain unchanged, the pattern of under- and overpressures due to topographically driven flow will not be subject to pressure equalization by flow. In fact, it is the flow itself that is responsible for the abnormal pressures.

A simple example of abnormal fluid pressures due to topographically driven flow is provided by considering flow between an alternating series of hills and valleys (Figure 8.6). The level to which water will rise in a cased well open at the bottom is determined by fluid head at the bottom of the well. This is equal to the elevation of the water table at the point the head contour at that depth intersects the water table. Note that in areas of descending flow, head contours are bent concave downwards. As a result, the head contours that intersect the bottom of the well will not rise to the top of the well, and the water level in the well will be depressed below the top of the well. In the discharge area at the bottom of the hill, the opposite situation prevails. Head contours are bent convex upwards, and water levels in wells located here will rise above the top of the wells and artesian flow will occur.

What is the fluid pressure (P, kg-m^{-1}-s^{-2}) at the bottom of each well? It is simply equal to the product of fluid density (ρ, kg-m^{-3}), acceleration due to gravity (g, m-s^{-2}), and height of the fluid column (z, m) in the well ($P = \rho g z$, equation 2.23). Thus, in areas of high elevation and descending flow where the fluid level falls below the water table, the fluid pressures at depth are below hydrostatic. These regions are therefore *underpressured.* Conversely, wells in the valleys where flow is ascending are *overpressured.* Thus fluid is moving from regions of underpressures to regions of overpressures! This is a striking example of the importance of using head to characterize flow regimes instead of pressure. Consider, for example, the disastrous consequences of placing a hazardous waste dump in an underpressured area with active flow.

8.3.2. Transient Mechanisms.

Following Neuzil (1995), we can rewrite the diffusion equation (equation 5.35) as

$$K\nabla^2 h = S_s \frac{dh}{dt} + \Gamma \qquad \textbf{(8.1)}$$

where K (m-s^{-1}) is hydraulic conductivity, h (m) is head, ∇^2 is the Laplacian (m^{-2}), S_s is the specific storage (m^{-1}), t (s) is time, and Γ (s^{-1}) is a geologic forcing term that represents the geologic agent responsible for abnormal pressure generation. In general, transient processes responsible for

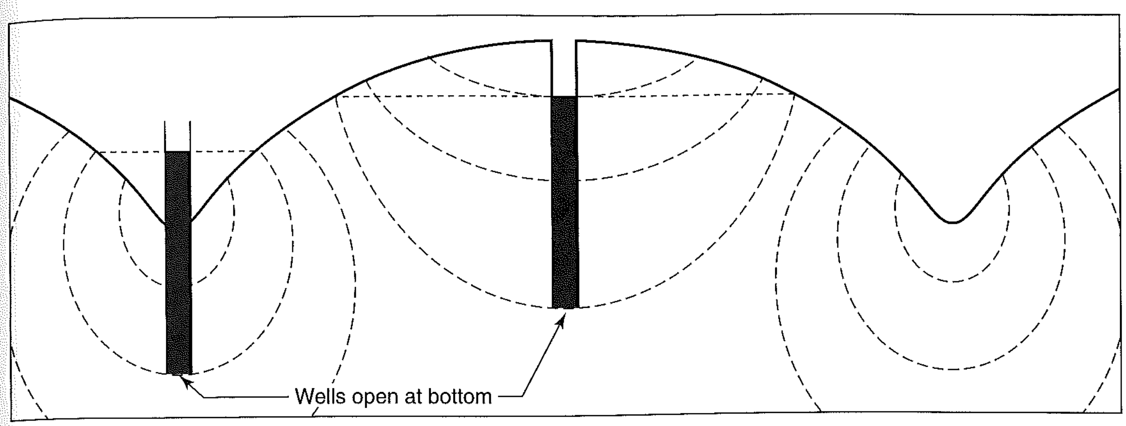


Figure 8.6 Cross-sectional plot of head (dashed lines) underneath a hill symmetrically flanked by valleys. Wells are present at the apex of the hill and the bottom of the valley as shown. The wells are open only at the bottom, so that the water level in each well is determined by head at the bottom of each well. Water rises in each well to the point at which head contours at the bottom of the wells intersect the water table at the surface. Fluid regime in recharge areas is underpressured; fluid regime in discharge areas is overpressured. Fluid thus flows from regions of underpressures to regions of overpressure.
(After Hubbert, 1940.)

abnormal pressure generation can be divided into three categories: (1) thermomechanical response of the fluid and matrix, (2) porosity changes due to stress changes and diagenesis/metamorphism, and (3) fluid sources and sinks (Neuzil, 1995, p. 748).

An example of abnormal pressures created by a thermal change in a fluid is aquathermal pressuring, that we considered earlier in section 7.1.3. Aquathermal pressuring is generally not as important as porosity changes, although the relative importance depends upon the local geothermal gradient and burial rate.

Underpressuring due to erosion is an example of abnormal pressure generation through the mechanical response of a porous medium to a change in stress (Toth and Millar, 1983). As erosion occurs, overburden weight decreases and the void spaces in a porous rock will tend to expand, just as they contracted in response to an increase in burial depth and effective stress. Generally speaking, a rock will not recover all of the porosity it had during its burial at the same depth. The failure to completely recover original porosity upon exhumation is an example of hysteresis. **Hysteresis** is the failure of a property that has been changed by an external agent to return to its original value when the cause of the change is removed. In this case, the external agent that causes a porous rock to compact is an increase in effective stress due to an increase in overburden weight from sedimentation. If and when the process reverses and the overburden is removed by erosion, porosity increases, but not all of the original porosity is recaptured.

Underpressuring can occur when the hydraulic diffusivity of a porous medium is so low that fluid flow cannot occur quickly enough to equalize head gradients created by pore space expansion. Pore space expansion is essentially synchronous with erosion, as stress is transmitted through rocks virtually instantaneously. The elastic response of the rock is also relatively fast, however, pressure equalization by fluid movement is a diffusive process whose rate is described by a characteristic time constant (equation 5.86) that is determined by the hydraulic diffusivity of a medium undergoing a change and its thickness.

Erosional unloading has been invoked to explain the existence of underpressures in Cretaceous shales of the Western Canadian Sedimentary Basin in Alberta, Canada (Parks and Toth, 1995).

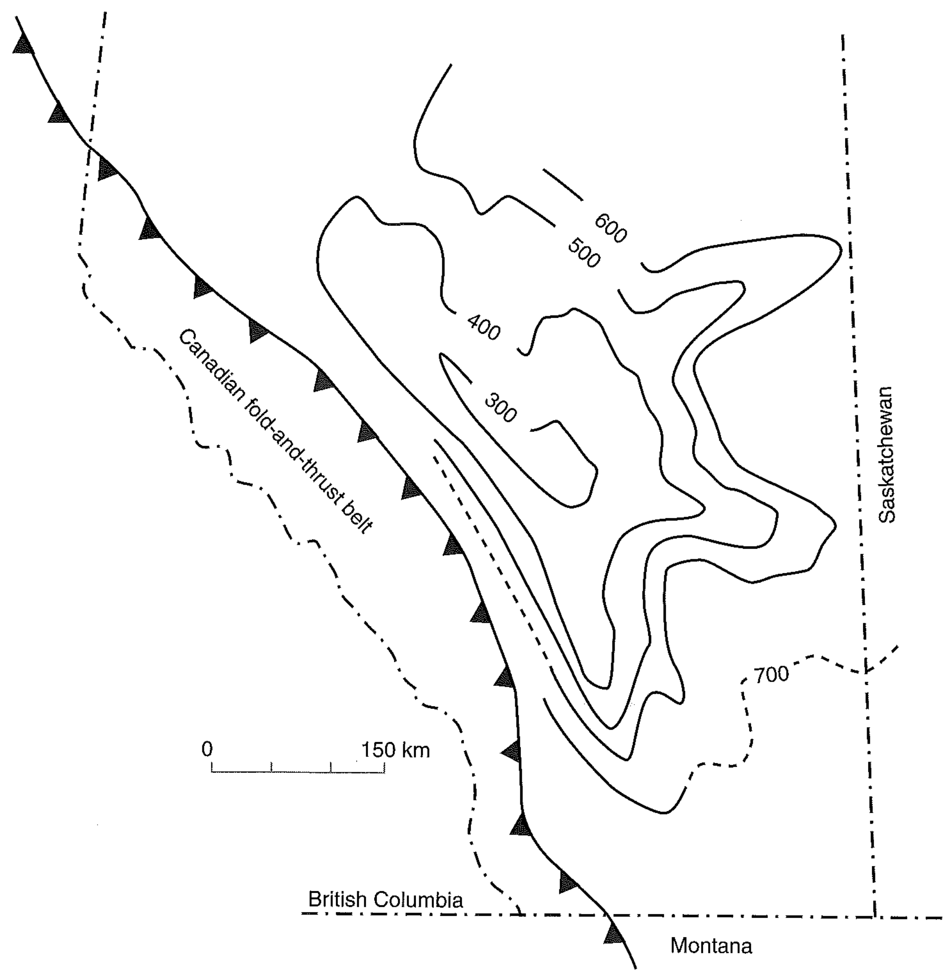


Figure 8.7 Fluid head (meters) in Lower Cretaceous rocks of the Western Canadian Sedimentary Basin. Closed minimum suggests the presence of a transient underpressure possibly related to pore expansion from erosion and unloading.
(From Corbet and Bethke, 1992, p. 7204, after Hitchon, 1969.)

The existence of head minima, which appear to be closed in three dimensions (Figure 8.7), suggests a disequilibrium condition that cannot be sustained over substantial periods of geologic time. Corbet and Bethke (1992) studied underpressures in this area and concluded that for the underpressures to be due to erosional unloading, shale permeabilities would have to be lower than 3×10^{-20} m^2. This is near the lower end of shale permeability, but not unusually low. In such situations it is difficult to show uniquely that a specific mechanism is responsible for an observed hydraulic phenomenon. The best that can usually be done is to delineate the circumstances under which a particular mechanism may operate, and then compare those findings with observational data.

Compaction disequilibrium is another example of a geologically important mechanism that is responsible for the existence of overpressures through porosity changes. Compaction disequilibrium occurs when the pore fluid sustains part of the matrix overburden weight, due to the failure

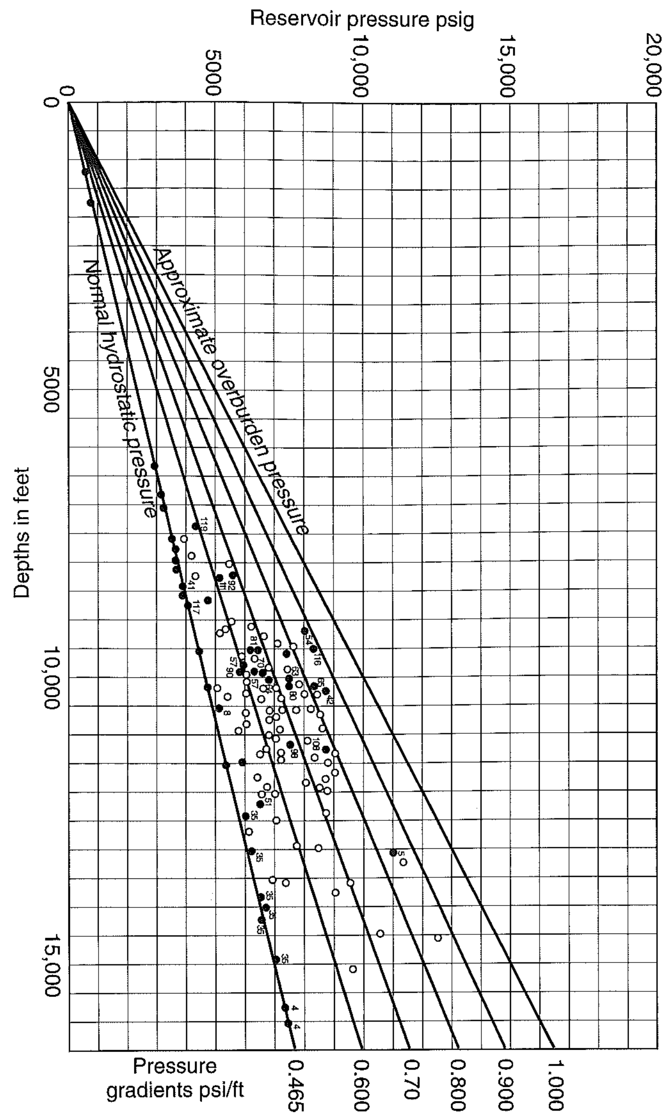


Figure 8.8 Reservoir fluid-pressure versus depth for Louisiana Gulf Coast wells. Solid circles correspond to measured pressures, open circles represent estimated fluid pressures.
(From Dickinson, 1953, 414.)

of sediments to reach equilibrium compaction conditions quickly enough. Porosity reduction is inhibited by the difficulty in expelling pore fluids from low-permeability shales and clay-rich sediments. The creation of overpressures by compaction disequilibrium requires high sedimentation rates and a predominance of low-permeability sediments or rocks.

Two examples of geopressures that are thought to be due to compaction disequilibrium are the Gulf Coast Basin of the southeast U.S. (Figure 8.8) and the South Caspian Basin of the former USSR (Bredehoeft and Hanshaw, 1968; Sharp and Domenico, 1976; Bethke, 1986; Bredehoeft et al., 1988; Mello et al., 1994). In the South Caspian Basin, sedimentation rates as high as ~1300 m-Ma^{-1} have resulted

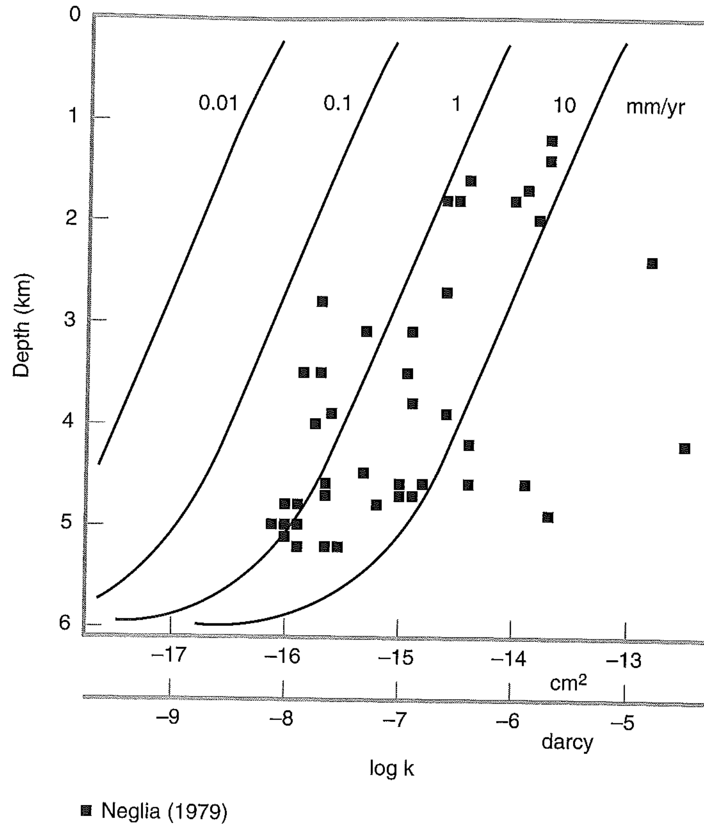


Figure 8.9 Permeability profiles required for overpressuring in a 6-km-deep sedimentary basin as a function of sedimentation rate in millimeters per year. Solid squares represent permeability measurements on sediments from the Gulf Coast basin as reported by Neglia (1979).
(From Bethke, 1986, p. 6538.)

in enormous thicknesses of sediments (25 km in deepest sections) and substantial overpressures (Bredehoeft et al., 1988). The Gulf Coast Basin in the southeast United States has been studied more intensively. Most rocks in the Gulf Coast Basin are shales overlain by deltaic systems containing alternating series of sands and shales grading vertically upward into massive sandstones (Mello, 1994, p. 2776). The sandstones are hydrostatically pressured; overpressures develop either in the mixed sands/shales, or in the underlying shale sequences. Overpressures result when pore fluids are not able to escape from low-permeability shales quickly enough to maintain a "normal" porosity-depth curve. The hydrological properties of the shales thus play a critical role in the maintenance and dissipation of overpressure in the Gulf Coast Basin, which contains more than 85% shale and shaley sediment (Bethke, 1986, p. 6539). For example, Bethke (1986, p. 6538) showed that anomalous formation pressures in the Gulf Coast basin could be maintained at sedimentation rates of 100-10,000 m-Ma^{-1} (0.1–10 millimeters-yr^{-1}) if average shale permeabilities were in the range of 10^{-18}-10^{-20} m^2. These values are consistent with laboratory measurements of shale permeabilities from the Gulf Coast (Neglia, 1979) as well as current sedimentation rates of 1 to 5 millimeters per year (Figure 8.9). Bethke (1986) also showed that deeper basins can develop overpressures in more permeable sediments than shallow basins (review section 5.5) (Figure 8.10). As overpressuring

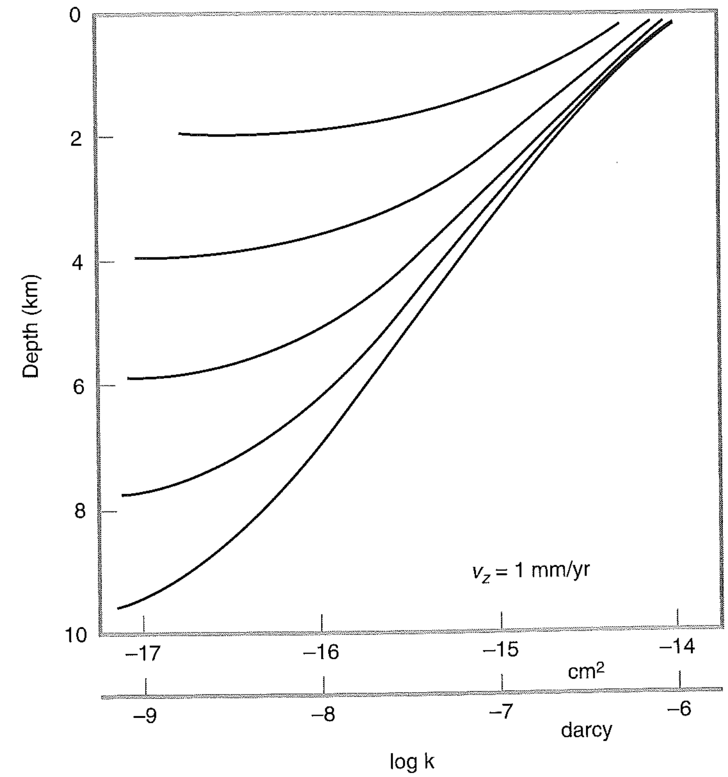


Figure 8.10 Permeability profiles required for overpressuring at a sedimentation rate of 1 millimeter per year as a function of total basin depth. Overpressuring can occur in deeper basins with higher average permeabilites. (From Bethke, 1986, p. 6537.)

from compaction disequilibrium is a transient phenomenon, this follows directly from the relationship between characteristic time and length (equation 5.86).

Despite the low permeability needed for the development of overpressures from compaction disequilibrium, there is evidence that large volumes of fluid have escaped from the Gulf Coast overpressured zone. Crude oils in the thermally immature Tertiary reservoirs of the shallow hydropressured zone are generally believed to be derived from Cretaceous or Jurassic source rocks within the geopressured zone (Nunn and Sassen, 1986; Kennicutt et al., 1992; Whelan et al., 1994). The presence of significant volumes (~10 percent) of secondary porosity and diagenetic cements in some Gulf Coast sandstones from the geopressured zone has also been interpreted as evidence for large volumes of fluid circulation within the geopressured zone itself.

At the present time, it is poorly understood exactly how fluids escape from, or move through, the overpressured zone in the Gulf Coast. Relatively low salinities in the overpressured zone tend to suggest that membrane filtration is not taking place, and thus uniform flow through pore spaces seems unlikely (see Figure 4.1 and discussion in section 4.3). Fluid may move laterally until it encounters a fault or fracture that enables it to escape into the overlying hydrostatically pressured zone. Nunn (1996) reviewed evidence that geopressured sediments in the Gulf Coast basin are mechanically weak, and suggested that upward fluid movement could be caused by buoyancy-driven propagation of isolated fluid-filled fractures. Nunn's (1996) calculations showed that isolated, fluid-filled fractures with lengths of a few meters or more can propagate through geopressured sediments with velocities of 1000 m-yr^{-1}.

The conversion of the clay mineral **smectite** to **illite,** another clay mineral, releases water, and may contribute to the development of overpressures in some sedimentary basins. Smectite is a common mineral found in shales and contains abundant water between the layers of its crystal structure. Under conditions of high pressure and temperature, the water is expelled from smectite and it is converted to illite. The precise nature of the diagenetic reactions involved and their exact dependence upon pressure and temperature conditions is poorly known. Some water may be expelled at temperatures lower than 60 °C; however, temperatures as high as 200 °C may be required for complete expulsion. The net volume change that takes place is also uncertain. Estimates range from about 4 to 40%. The primary evidence that implicates the smectite-illite conversion in overpressuring is an observed change in the ratio of smectite to illite that occurs near the top of the overpressured zone. This coincidence has been observed, for example, in the Gulf Coast Basin of the southeast United States (Bethke, 1986), however, geopressures are already well developed in the Caspian Basin, even though the smectite-illite ratio remains unchanged down to depths of 6 km. This counterexample shows that the conversion by itself cannot be responsible for the generation of overpressures in rapidly subsiding basins similar to the Caspian or Gulf Coast Basin. It may be that the smectite-illite conversion is related to the development of overpressures by reducing permeability, instead of acting as a fluid source.

Petroleum generation is probably the most theoretically significant fluid source capable of creating abnormal fluid pressures. Neuzil (1995) estimated the magnitude of Γ (the forcing function, eq. 8.1) for different geologic mechanisms, and concluded that the mechanism with the largest probable magnitude was petroleum generation. Neuzil (1995, p. 758) estimated that the magnitude of Γ for petroleum generation can be as large as 10^{-14} s^{-1}, while other mechanisms evaluated by Neuzil result in Γ estimates typically on the order

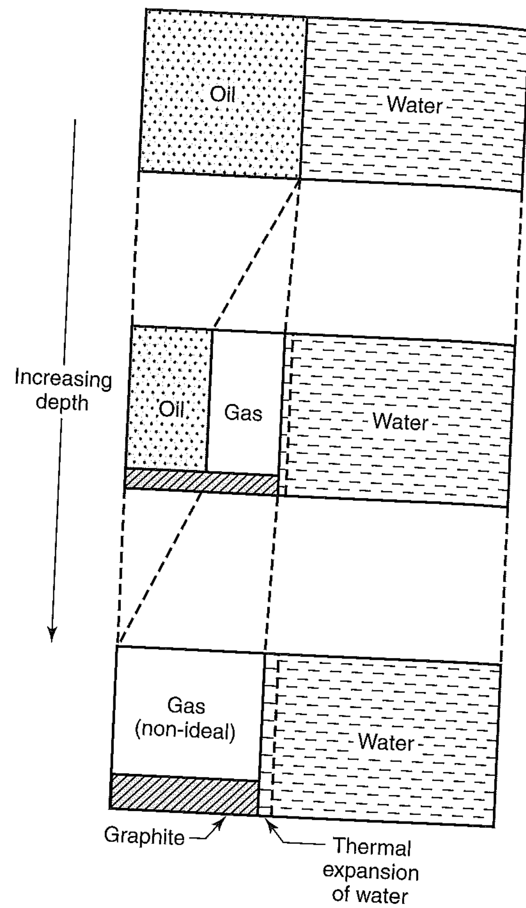


Figure 8.11 Conceptual model of liquid petroleum cracking to gas with increasing depth, time, and temperature. Cracking process also produces a graphite residue. Volume changes not shown to scale. (From Barker, 1990, p. 1257.)

of 10^{-15} s^{-1}. Other researchers (e.g., Barker, 1990), however, have shown that gas generation is a much more effective mechanism for overpressuring than oil generation. Overpressure from oil generation is a consequence of higher-density kerogen being replaced by lower-density oil, which requires more volume for the same mass. Kerogen is the solid organic material that breaks down to form oil and gas at high temperatures. As natural gas has a much lower density than liquid oil, it follows that gas generation is a more efficacious mechanism for overpressure generation than oil generation (Figure 8.11). Barker (1990) estimated that 85 to

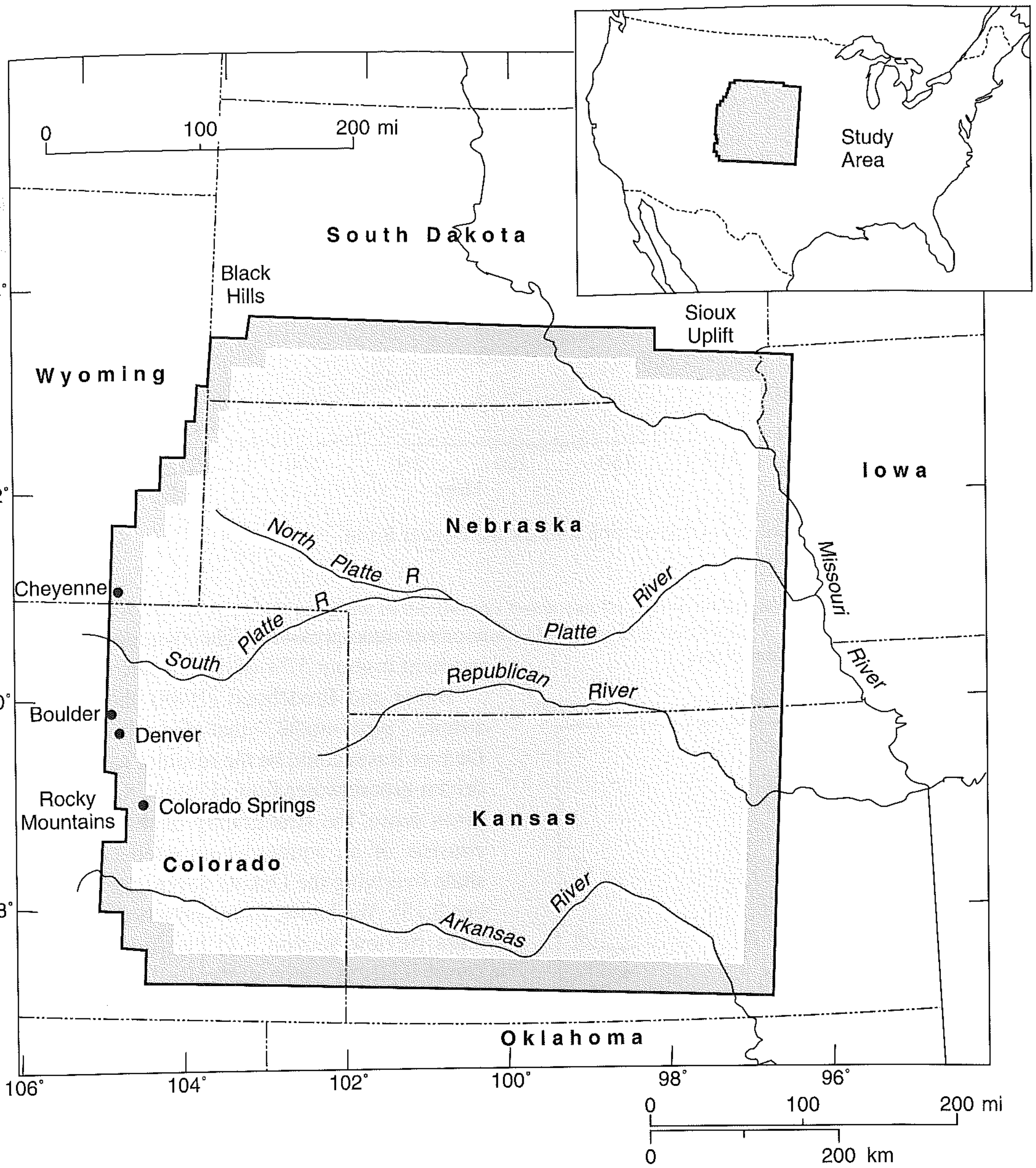


Figure 8.12 Location of the Denver Basin in the Central U.S.
(From Belitz and Bredehoeft, 1988, p. 1335.)

113 m^3 of gas is generated by each barrel (1 barrel = 42 gallons = 158.98 liters = 0.159 m^3) of oil that turns into gas at high temperatures. Lithostatic pressures can thus be reached after only 1% of the oil in a reservoir cracks into gas.

8.4. CASE STUDY: UNDERPRESSURES IN THE DENVER BASIN.

The Denver Basin in the central United States (Figures 8.12 and 8.13) is known to have extensive

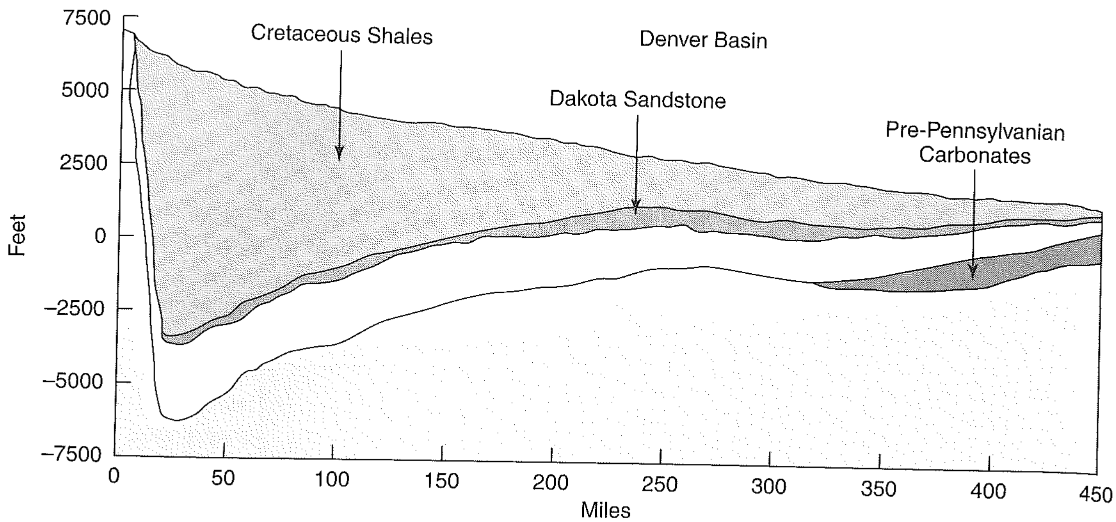


Figure 8.13 Generalized geologic cross-section through the Denver Basin.
(From Belitz and Bredehoeft, 1990.)

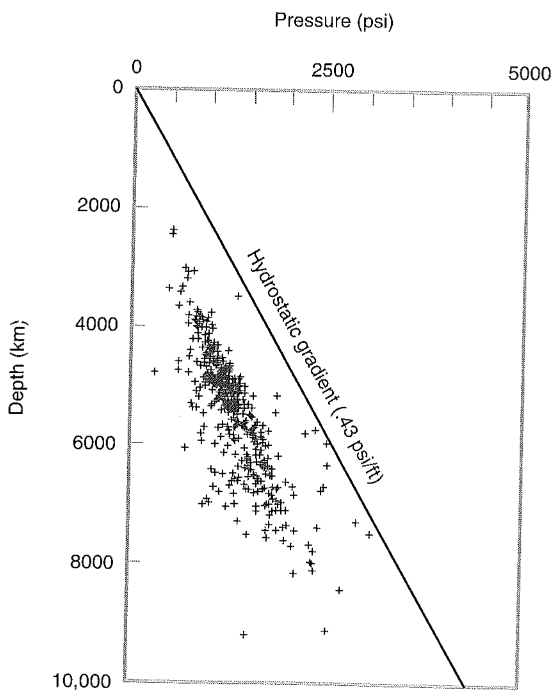


Figure 8.14 Fluid pressure versus depth for Denver Basin sandstones.
(From Belitz and Bredehoeft, 1988, p. 1335.)

areas of underpressures. The average fluid pressure gradient is about 57% of hydrostatic (Figure 8.14).

Belitz and Bredehoeft (1988) showed, using a simple "pipe model," that underpressures in the Denver Basin could be the result of topographically driven groundwater flow (Figures 8.15, 8.16). In a gross sense, the stratigraphy of the Denver Basin consists of an aquitard composed of Cretaceous shale overlying the Dakota Sandstone aquifer (Figure 8.15). We know, from refraction of head contours (review section 5.4) that flow in confining layers tends to move vertically. Thus, flow in the recharge region must be nearly vertical downwards through the Cretaceous shales into the Dakota Sandstone. In the Dakota Sandstone the direction of flow must be nearly horizontal, parallel to bedding. It is thus possible to analyze flow in the Denver Basin with a simple "pipe model" (Figure 8.16). The vertical pipe represents flow through the Cretaceous Shale aquitard; the nearly horizontal pipe represents flow through the Dakota Sandstone aquifer. At the top of the vertical pipe, head at the recharge site (h_r, m) is fixed by elevation z_r (m) (we assume that the water table is at the ground surface). The length of the vertical pipe is equal to the depth of the basin, D (m). The hydraulic conductivity of the vertical pipe

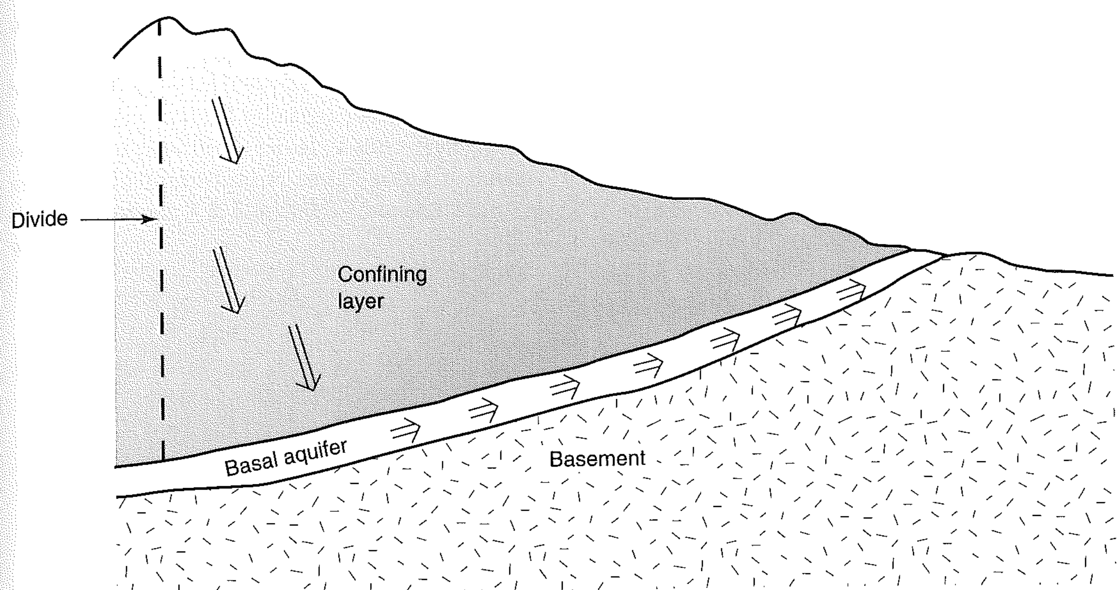


Figure 8.15 Topographically-driven flow in the Denver basin can be approximated by near-vertical flow through a Cretaceous Shale confining layer into a basal aquifer where flow is nearly horizontal.

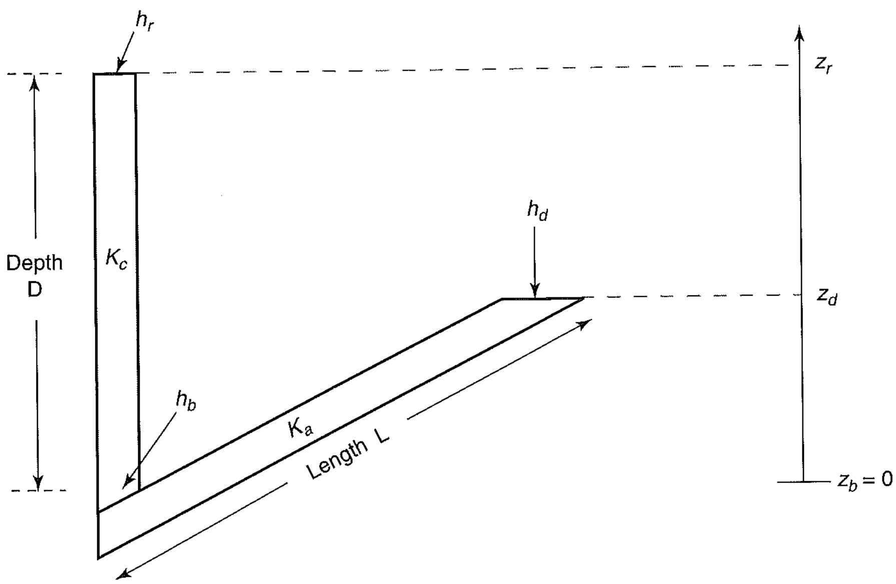


Figure 8.16 Pipe model for topographically-driven flow through the Denver Basin.
(After Belitz and Bredehoeft, 1988.)

representing the confining layer is K_c (m-s^{-1}). The horizontal pipe has length L (m), equal to the length of the basin. Head is fixed at the end of the horizontal pipe in the discharge region (h_d, m) by elevation (we again assume the water table is at the ground surface). The horizontal pipe representing the

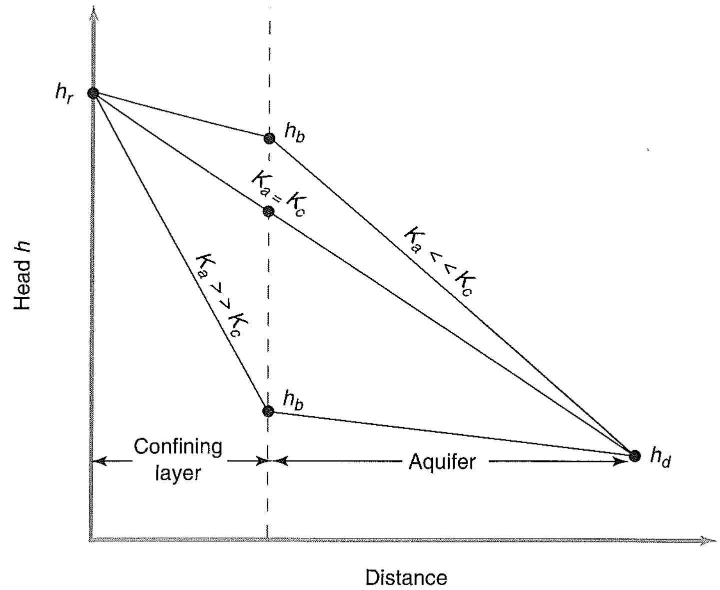


Figure 8.17 Schematic illustration of head gradients for pipe-model scenarios with contrasting hydraulic conductivities.

aquifer has hydraulic conductivity K_a (m-s^{-1}). Head at the boundary between the two pipes (h_b, m) is unknown. For simplicity, we will define an arbitrary datum for head and elevation by defining the elevation of the boundary between the vertical and horizontal pipes to be zero. Note that both h_r and h_d are greater than zero.

If we assume conservation of mass and steady-state flow, then the Darcy velocity in the confining layer (q_c, m-s^{-1}) must be equal to the Darcy velocity in the aquifer (q_a, m-s^{-1}). Applying Darcy's Law, we obtain

$$\frac{K_c(h_r - h_b)}{D} = \frac{K_a(h_b - h_d)}{L} \tag{8.2}$$

or, rearranging equation 8.2,

$$\frac{K_c}{K_a}\frac{L}{D} = \frac{(h_b - h_d)}{(h_r - h_b)} \tag{8.3}$$

Equation 8.3 implies that the value of head at the boundary between the confining layer and the basal aquifer (h_b) is determined by the geometry of the basin (the ratio L/D) and the ratio of the hydraulic conductivities (K_c/K_a) (Figure 8.17). If $K_a >> K_c$, the head gradient in the confining layer is much greater than in the aquifer. Consequently, h_b is relatively low. Conversely, if $K_c >> K_a$, the head gradient in the confining layer would be much lower than in the aquifer, and h_b would be relatively high. Of course, in the latter situation, the designations "confining layer" and "aquifer" would have to be reversed.

Recall that head (h, m) has an elevation and pressure component (equation 2.45)

$$h = z + \frac{P}{\rho g} \tag{8.4}$$

where z is elevation (m), P is fluid pressure (kg-m^{-1}-s^{-2}), ρ is fluid density (kg-m^{-3}), and g is the acceleration due to gravity (m-s^{-2}). Thus,

$$h_b = 0 + \frac{P_b}{\rho g} \tag{8.5}$$

$$h_d = z_d + 0 \tag{8.6}$$

$$h_r = D + 0 \tag{8.7}$$

where the subscript b indicates the boundary between the vertical and horizontal pipes, and we have specified fluid pressure (P) in terms of gauge pressure. Thus, at the recharge and discharge sites where fluid pressure is equal to atmospheric pressure, the fluid pressures P_r and P_d are zero. Substituting equations 8.5, 8.6, and 8.7 into equation 8.3 we obtain

$$\frac{K_c}{K_a}\frac{L}{D} = \frac{\dfrac{P_b}{\rho g} - z_d}{D - \dfrac{P_b}{\rho g}} \tag{8.8}$$

$$\left[D - \frac{P_b}{\rho g}\right]\frac{K_c}{K_a}\frac{L}{D} = \frac{P_b}{\rho g} - z_d \tag{8.9}$$

$$D\frac{K_c}{K_a}\frac{L}{D} - \frac{P_b}{\rho g}\frac{K_c}{K_a}\frac{L}{D} - \frac{P_b}{\rho g} = -z_d \tag{8.10}$$

$$\frac{P_b}{\rho g}\left[\frac{K_c}{K_a}\frac{L}{D} + 1\right] = \frac{K_c}{K_a}L + z_d \tag{8.11}$$

$$\frac{P_b}{\rho g} = \frac{\dfrac{K_c}{K_a}L + z_d}{\dfrac{K_c}{K_a}\dfrac{L}{D} + 1} \tag{8.12}$$

$$P_b = \rho g \left\{\frac{\dfrac{K_c}{K_a}L + z_d}{\dfrac{K_c}{K_a}\dfrac{L}{D} + 1}\right\} \tag{8.13}$$

$$P_b = \rho g D \left\{\frac{\dfrac{K_c}{K_a}L + z_d}{\dfrac{K_c}{K_a}L + D}\right\} \tag{8.14}$$

Equation 8.14 gives the fluid pressure at the boundary between the confining layers and aquifer in terms of the hydrostatic fluid pressure ($\rho g D$) and a term (in brackets) that depends upon the hydraulic conductivity of the confining layer and aquifer as well as the basin geometry (length and depth).

Note that the basin is underpressured even if $K_c = K_a$. In this case, the ratio of the pressure gradient in the basin to a hydrostatic gradient is

$$\frac{L + z_d}{L + D} \tag{8.15}$$

If we take $L = 644$ km (400 miles), $z_d = 1.676$ km (5500 ft), and $D = 3.048$ km (10,000 ft) for the Denver Basin (see Figure 8.13), the fluid pressure gradient is 99.8% of hydrostatic.

Problem: What must the ratio K_c/K_a be for topographically driven flow in the Denver Basin to result in the fluid pressures being 57% of hydrostatic?

Denote the dimensionless ratio of fluid pressure (P_b at depth D) to hydrostatic fluid pressure ($\rho g D$) as γ. From equation 8.14,

$$\frac{P_b}{\rho g D} = \gamma = \frac{\dfrac{K_c}{K_a}L + z_d}{\dfrac{K_c}{K_a}L + D} \tag{8.16}$$

Now solve for the ratio K_c/K_a in terms of γ. Rearranging equation 8.16,

$$\frac{K_c}{K_a}L + z_d = \gamma\left[\frac{K_c}{K_a}L + D\right] \tag{8.17}$$

Gathering terms,

$$\frac{K_c}{K_a}L(1 - \gamma) = \gamma D - z_d \tag{8.18}$$

$$\frac{K_c}{K_a} = \frac{\gamma D - z_d}{L(1 - \gamma)} \tag{8.19}$$

For $\gamma = 0.57$ as observed, $D = 3.048$ km, $z_d = 1.676$ km, and $L = 644$ km,

$$\frac{K_c}{K_a} = 2.2 \times 10^{-4} \tag{8.20}$$

Equation 8.20 implies that the aquifer is about 4,500 times more permeable than the overlying confining layer.

Review Questions

1. Define the following terms in the context of hydrogeology:
   a. hydrostatic
   b. lithostatic
   c. underpressures
   d. overpressures
   e. geopressures
   f. static school
   g. pressure seal
   h. pressure compartment
   i. capillary force
   j. dynamic school
   k. hysteresis
   l. compaction disequilibrium
   m. smectite
   n. illite
   o. kerogen
2. Which is more common in sedimentary basins—overpressures or underpressures?
3. If a sedimentary basin is overpressured, at what depths are overpressures usually first encountered?
4. What are the two schools of thought on the creation and preservation of abnormal pressures in the Earth's crust?
5. What are some of the difficulties with the pressure seal concept?
6. What are the advantages of hypothesizing gas capillary forces as pressure seals?
7. Use a scale analysis to estimate the minimum permeability necessary for a layer 50 m thick layer to confine abnormal fluid pressures for 10 Ma (review section 5.5).
8. Show how overpressures and underpressures may both be due to steady-state topographically driven flow. Use a figure. Explain how fluid can be flowing from areas where fluid pressure is below hydrostatic to areas where fluid pressure is above hydrostatic.
9. Is a good idea to locate a toxic waste dump in an underpressured area? Why or why not?
10. What are two geologic processes that can lead to underpressuring?
11. What process is believed to be responsible for the existence of overpressures in the Gulf Coast Basin of the southeast U.S.? What geologic factors have contributed to the development of geopressures in this area?
12. In what two ways could the conversion of smectite to illite contribute to overpressuring?
13. Explain how oil and/or gas generation can lead to overpressuring? Which (oil or gas generation) is more likely to lead to overpressuring? Why?
14. What is the *maximum* underpressuring that can occur in the Denver Basin?

Suggested Reading

Belitz, K., and Bredehoeft, J. D. 1988. Hydrodynamics of Denver Basin. Explanation of subnormal fluid pressures. *American Association of Petroleum Geologists Bulletin,* 72: 1334–1359.

Bethke, C. M. 1986. Inverse hydrologic analysis of the distribution and origin of Gulf Coast-type geopressured zones. *Journal of Geophysical Research,* 91: 6535–6545.

Osborne, M. J., and Swarbrick, R. E. 1997. Mechanisms for generating overpressures in sedimentary basins: a reevaluation. *AAPG Bulletin,* 81: 1023–1041.

Notation Used in Chapter Eight

Symbol	Quantity Represented	Physical Units
D	basin depth	m
Γ	forcing term for geologic overpressuring	s^{-1}
g	acceleration due to gravity	$m\text{-}s^{-2}$
γ	ratio of fluid pressure to hydrosatic	dimensionless
h	head	m
h_b	head at boundary between confining layer and aquifer	m
h_d	head in discharge region	m
h_r	head in recharge region	m
K	hydraulic conductivity	$m\text{-}s^{-1}$
K_a	hydraulic conductivity of basin aquifer	$m\text{-}s^{-1}$
K_c	hydraulic conductivity of confining layer	$m\text{-}s^{-1}$
L	basin length	m
P	fluid pressure	Pascal (Pa) = $kg\text{-}m^{-1}\text{-}s^{-2}$
P_b	fluid pressure at boundary between confining layer and aquifer	Pascal (Pa) = $kg\text{-}m^{-1}\text{-}s^{-2}$
P_d	fluid pressure of discharge region	Pascal (Pa) = $kg\text{-}m^{-1}\text{-}s^{-2}$
P_r	fluid pressure of recharge region	Pascal (Pa) = $kg\text{-}m^{-1}\text{-}s^{-2}$
ρ	fluid density	$kg\text{-}m^{-3}$
S_s	specific storage	m^{-1}
z	elevation	m
z_d	elevation of discharge region	m
z_r	elevation of recharge region	m
∇^2	Laplacian, the second spatial derivative	m^{-2}