More than a seal - the impact of shales on the flow and evolution of formation water

Karsten Michael

CSIRO Earth Science & Resource Engineering, 26 Dick Perry Avenue, Kensington WA 6151

Abstract

Shales generally have a low permeability and are barriers to fluid flow, thereby separating flow systems in various aquifers or forming seals for hydrocarbon accumulations. However, shales may also contribute to driving fluid flow and to the chemical evolution of formation water. Low-permeability environments in subsiding basins are prone to overpressuring due to disequilibrium compaction. In contrast, when a basin is uplifted underpressures may develop due to the erosional rebound of the sediments. Shales are also thought to act as semi-permeable membranes, which may locally drive vertical flow and lead to salinity changes. Diagenetic transformations of clay minerals can also cause overpressuring for example due to additional water from smectite dehydration or reduction in pore space as a result of the precipitation of cements due to the release of ions. Quantification of the resulting fluid fluxes from most of the aforementioned processes is difficult because of the long time-scales involved and uncertainty in changes of hydraulic parameters and boundary conditions during that time.

Introduction

In contrast to near-surface groundwater flow, which is almost solely driven by gravity, regional flow in sedimentary basins is driven by a combination of various mechanisms such as: a) topography (gravity), b) sediment compaction, c) tectonic loading, d) erosional rebound, e) buoyancy, f) overpressures due to hydrocarbon generation, and g) osmosis or mineral phase changes. The latter has only a small effect on fluid flow, but contributes to the change of formation water chemistry. During the geological history of a sedimentary basin shales may have various impacts on driving mechanisms and fluid chemistry, which are the topic of this review.

Flow-driving mechanisms

It is generally accepted that one of the most important driving mechanisms for fluid flow at various scales in continental sedimentary basins is gravity, and that flow is controlled by the ground surface. Topography-driven flow may be active almost throughout the entire basin history (Fig. 1a-e). At least for shallow to intermediate depths, the potentiometric surface of hydraulic heads is generally a smoothed replica of the ground surface. This indicates that there is groundwater recharge in areas of topographic highs and water discharging in areas of topographic lows. With increasing depth the similarity between potentiometric surfaces of various aquifers and topographic relief decreases, which results in local, intermediate, and regional flow systems (Tóth, 1963). The architecture of the basin and the permeability distribution control the extent of the various flow systems. Thick, regionally extensive aquitards, largely formed by shales, impede cross-formational flow and separate different flow systems above and below. Other flow-driving mechanisms become significant only in low-permeability environments and/or at greater depths with increasing temperature and salinity. They usually become important during certain stages of basin evolution.

Compaction and erosion

An important driving mechanism in the early stages of basin evolution is compaction due to burial of the sediments and pore fluids (Fig. 1a). Compaction is a process in which, due to the weight of overburden, porosity decreases, while pressure and temperature increase, and fluids are expelled from the reduced pore space. The increased load on the sediment column is partitioned between the mineral grains (effective stress) and the pore fluid (fluid pressure). “Normal compaction” or “compaction equilibrium” implies concurrent fluid expulsion with loading of the sediments, without apparent increase in pore pressure, in contrast to “disequilibrium compaction”, where fluids are expelled some time after the deformation process, resulting in above-hydrostatic fluid pressures (Domenico & Schwartz, 1998). Whether normal or disequilibrium compaction prevails depends on sedimentation rate and permeability, which govern the drainage rate of the pore fluid, sediment compressibility, and rate of compaction. For normal compaction, the velocity of upward-moving pore water, controlled by the decrease in porosity due to newly added sediment, decreases from top to bottom of the sedimentary sequence and is proportional to the sedimentation rate (Einsele, 1978). In the case of a very high sedimentation rate in fine-grained sediments, the permeability is relatively low. Fluids cannot drain from the reduced pore space fast enough to let fluid pressures equilibrate to the increase in stress, which leads to overpressures in the fluid phase.

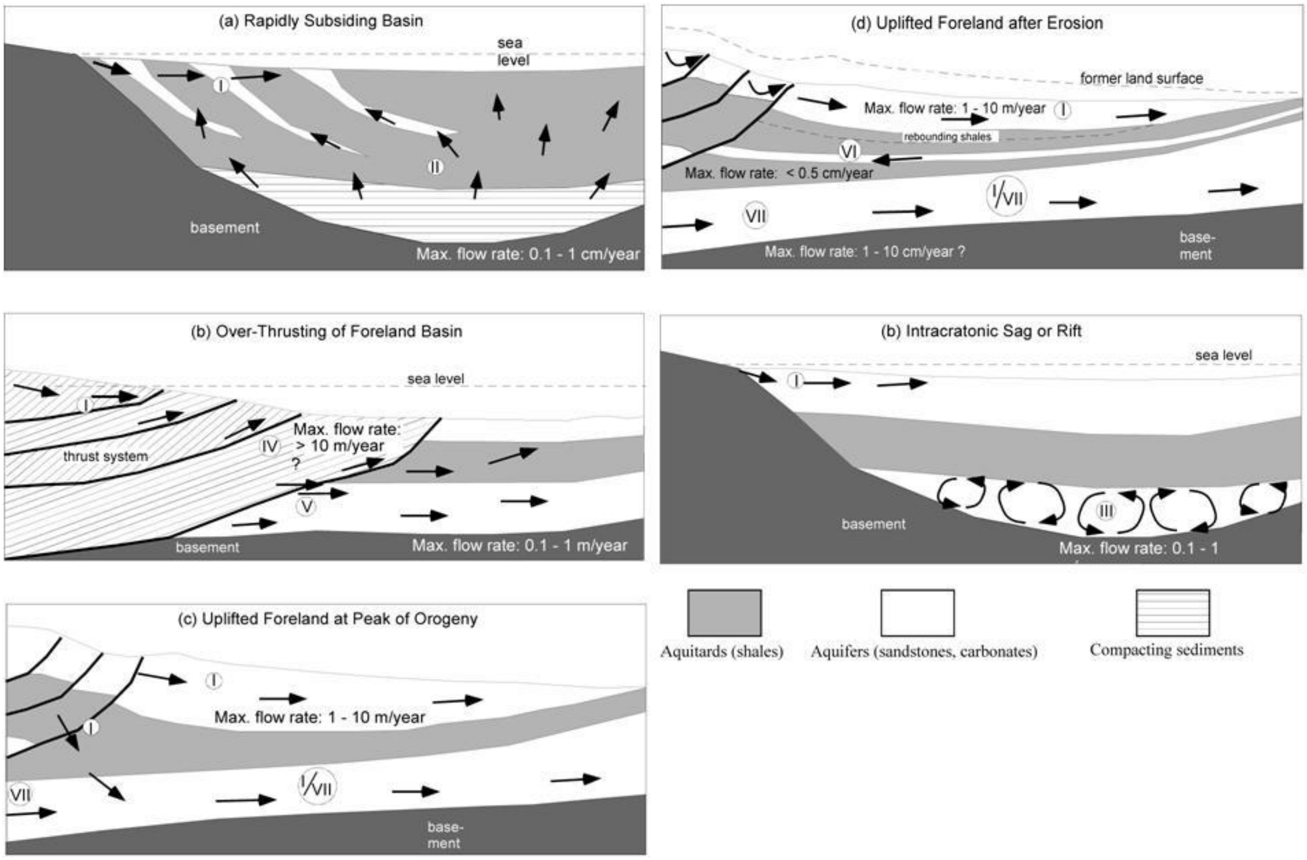

Figure 1. Conceptual representation of regional-scale hydrogeological regimes during the history of a sedimentary basin (modified from Garven, 1995). Arrows represent general flow direction. I Topography-driven flow; II Compaction-driven flow; III Thermally-driven free convection; IV Seismic pumping; V Tectonic expulsion; VI Flow driven by erosional rebound; VII Paleo-flow regime of tectonic origin or paleo-topography.

Erosional unloading and uplift of the basin at the end and after an orogeny cause a decrease in total stress. Neuzil & Pollock (1983) suggest that this may cause formerly compressed sediments to partly rebound, causing a dilatation of the rock framework, an increase in pore space and, concurrently, a decrease in fluid pressure (Fig. 1d). According to these authors, the sediment must have a very low permeability and a large compressive storage in order to prevent fluid pressures from equilibrating and to maintain sub-hydrostatic fluid pressures over a certain period of time. The best examples are thick, de-compacting shale units like the Cretaceous Pierre Shales in the Williston Basin (Neuzil & Pollock, 1983; Neuzil, 1993), and the Colorado shales in the Alberta Basin (Corbet & Bethke, 1992; Bachu & Underschultz, 1995; Parks & Tóth, 1995). Furthermore, the sediments should not be fractured or inter-bedded with coarser grained layers on a regional scale to an extent that allows pressures to equilibrate via connected high permeability pathways.

It is important to note that the specific storage value for compaction is larger than that for erosional rebound, because only a part of the compressive deformation of the rock framework is reversible (mechanical hysteresis). This implies that compaction has a much larger effect on fluid pressures than erosional rebound. Overall, it is relatively hard to quantify the effects of compaction, and especially those of erosional rebound, on fluid flow, because the change of many variables over time is not known. Permeability in particular changes significantly during these processes, and compaction and erosion rates are not constant, neither spatially nor over time.

Water-rock interaction

Mineral phase change reactions during increasing burial, such as gypsum to anhydrite and the transformation of clay minerals, result in a fluid volume increase due to the release of water. In the case of dehydration of gypsum to anhydrite, this volume increase is 38% compared to a 5 - 10% increase for the transformation of smectite to illite or the dehydration of montmorillonite (Bjørlykke, 1993). If fluids cannot escape the pore space fast enough, or the porosity is not increased by the expansion of the rock framework, the volume increase will lead to overpressures. According to Osborne & Swarbrick (1997), both processes are probably not responsible for significant overpressures, because the gypsum - anhydrite reaction mainly occurs during shallow burial and therefore cannot produce overpressures at greater depth. The dehydration of smectite is only responsible for a small release of water and is inhibited by a pressure buildup. A more effective source for overpressures seems to be the precipitation of cements due to the release of ions during clay transformations, which results in decreases in porosity and permeability (Osborne & Swarbrick, 1997). A much larger volume increase in the fluid phase is generated by the phase changes from kerogen to petroleum, further from kerogen degradation and thermal cracking of liquid to gaseous hydrocarbons, and direct gas generation in shales (Hedberg, 1974; Spencer, 1987).

Another process believed to drive fluid flow is chemical osmosis. Osmosis occurs when shales, which act as a semi-permeable membrane, separate two aqueous solutions of different salinity. The passage of anions and cations through the membrane is inhibited because of the positively and negatively charged layers of the shales. To equilibrate the chemical gradient across a membrane, neutral water molecules pass through the shale layer, resulting in a net flow of water to the more saline side of the membrane. Marine & Fritz (1981) explain pressures in excess of hydrostatic of about 100 kPa, or 10 m excess hydraulic head, in a Triassic basin in South Carolina by osmosis, which causes freshwater to pass through 30 - 300 m thick illite - rich shales and dilute the connate Triassic brines. Describing laboratory hydraulic tests on kaolinite, Olsen (1985) proposes that osmosis is responsible for non-Darcyian flow in low-permeability sediments. Like all the other mechanisms involving low-permeability environments, osmosis requires compacted, continuous, and non-fractured shale layers. However, according to Phillips (1983), shales are inefficient semi-permeable membranes. This, coupled with $K_C << K_H$, renders osmosis as a negligible driving mechanism for regional-scale flow (Bachu, 1995).

Conclusions

Shales play an important role for fluid flow in sedimentary basins because they:

- form major aquitards and form barriers between various aquifers and flow systems
- form seals for hydrocarbon accumulations
- provide a low-permeability environment that can sustain anomalous pressures
- may generate hydrocarbons
- may form semi-permeable membranes
- can produce excess freshwater from clay dehydration during burial

Quantification of the resulting fluid fluxes from many of the aforementioned processes is difficult because of the long time-scales involved and uncertainty in changes of hydraulic parameters and boundary conditions during that time.

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