DOMES and BASINS

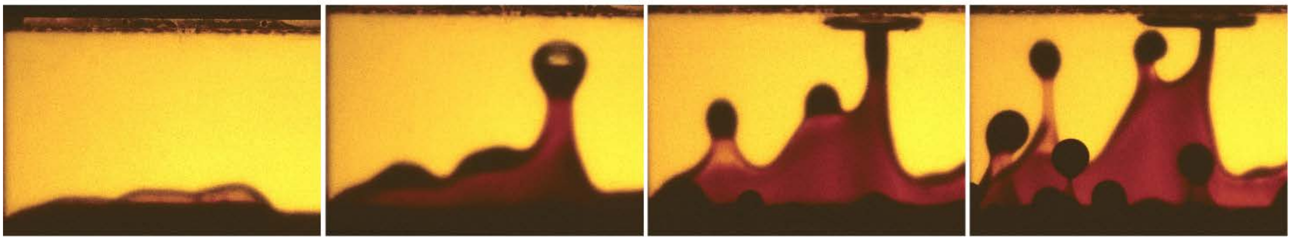
Domes and **basins** are structures with approximately circular or slightly elongate, closed outcrop patterns. Domes are convex upward; basins are concave upward.
Dome and basins have several origins.
*Compressional domes and basins: In many orogenic belts the motive force is ascribed to the compression responsible for associated folding. Main interpretations are:

- Folding of a basement-cover unconformity by two fold sets with different trends producing domes at the culminations of crossing anticlines. However, this mechanism should not produce equidimensional patterns everywhere.
- Reactivation of basement plutons: The development of the reference Finnish domes is a process requiring two orogenic events. During the first orogeny, granite plutons were emplaced in metasediments and metavolcanites that were then eroded to expose the plutons. The plutons and country rock were then covered by a new sequence of sediments. Finally, during the second orogeny, the old plutons were reactivated by injection of new granitic magma causing them to expand upwards and thereby deform the overlying strata into a dome. The new magma and deformation converted the old granite into migmatites and gneisses and gave rise to the small igneous bodies seen to intrude the cover around the gneiss domes.

*Gravity driven domes and basins: In many non-orogenic areas, domes and basins have resulted from the active ascent of mobile rocks. Such domes and basins resemble structures that develop spontaneously in a system where a heavy fluid is laid above a lighter fluid. The geometrical analogy has led to the notion of diapirs. Diapirs are closed, usually antiformal structures cored by relatively low-density rocks (evaporites, magma, etc.) that have locally broken through the crests or penetrated along faults the surrounding rocks.

- Diapirism: A common explanation postulates that the granite core rose in response to the gravity instability due to the low density of the granitic magma relative to the denser metamorphic rocks. This mechanism has been modelled mathematically and experimentally.
- Igneous intrusions: The core-mantle contact is intrusive and the dome shape is the primary igneous form.

Diapirism of the buoyant, dark fluid (oil) through a higher density fluid (glucose)
Unpublished experiment



*Extensional domes and basins: In regions of important continental extension, broad, fault bounded culminations represent horsts between sedimentary basins, equivalent to grabens. Gneiss and granitic intrusions in these culminations form **metamorphic core complexes**.

- Metamorphic core complexes: Extreme extension locally striped off the shallower layers to expose metamorphic and plutonic rocks that originally were below the detachment faults, at depths as great as 20 km.

Gravity-driven and compressional / extensional mechanisms are not mutually exclusive. Most gneiss domes have probably involved some buoyant diapirism since the density contrast is invariably present between the plutonic and anatectic components of the core rocks and surrounding rocks. Fold interference is not incompatible with the gravity instability mechanism since both mechanisms may operate synchronously.

Large-scale dome-and-basin structures

Large-scale dome-and-basin structures are often interpreted in terms of closed, type 1 (egg-box) interference patterns between two directions of upright folds. These interfering folds tend to be concentric or parallel in style with very little internal deformation of beds: for example pelitic beds show no foliation and sandstone layers are only strongly jointed. Slickensides are commonly developed on bedding planes indicating that flexural slip is an important part of the deformation mechanism. The lack of deformation within beds on the granular scale means that such dome-and-basin areas correspond to regions of relatively low overall shortening. The fact that associated folds are essentially parallel in style means that there are volume problems upward and downward along axial planes.

Gneiss domes

Gneiss domes are common in highly metamorphosed internal zones of orogenic belts. Within these zones, elongate groups of several domes often form a cluster-ridge. Gneiss domes commonly display foliation parallel to their periphery and are surrounded by metamorphosed or unmetamorphed, dominantly sedimentary rocks.

Mantled gneiss domes

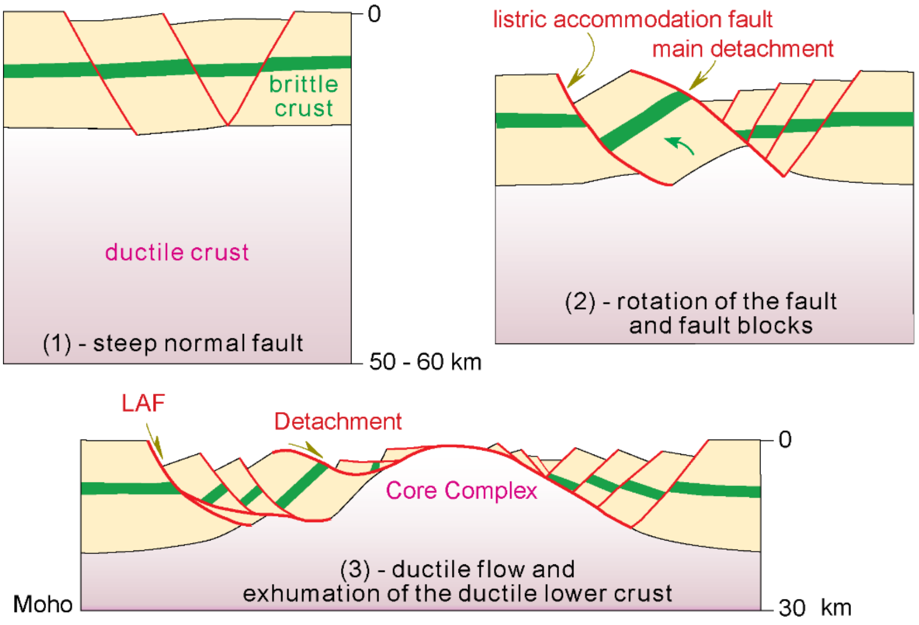
Mantled gneiss domes typically comprise a dome-shaped core of granitoids, migmatites and gneisses surrounded and overlain by a metasedimentary and metavolcanic cover awkwardly called **mantle,** since they actually are crustal rocks. Core rocks are almost invariably foliated near the margins. This foliation, the contact and the layering in the surrounding metamorphic cover are parallel and usually dip outward, away from the core. Regional, amphibolite facies metamorphism has commonly affected both the core and “mantle” rocks. The basal metasedimentary rock is generally a consistent stratigraphic horizon, in many cases a conglomerate with pebbles of the core granitoids and gneisses. Such cases suggest that the structure is a folded unconformity between cover and basement. Radiochronologic ages of core rocks older than surrounding rocks have often supported this interpretation. The Archean granite-greenstone terranes resemble large-scale mantled gneiss domes and may have formed by similar mechanisms.

Metamorphic Core Complexes

The Basin and Range Province in the western U.S.A. and northern Mexico is a continental system of regionally distributed normal faults-that continues into southwestern Canada. The Province has been extended by a bulk factor of two. Extension produced nearly 150 linear and parallel horsts forming mountain ranges of up to 4000 m height, alternating with grabens such as the Death Valley, that are mostly filled by the erosion products from the adjacent ranges. The little to unmetamorphosed cover sedimentary beds are highly rotated and attenuated by movement along many listric faults, cut off at depth (or merging with) shallow-dipping detachments. The cover sequences and underlying metamorphic core are typically separated by a mylonitic to cataclastic zone of generally upwarped main detachment that delineates a steep metamorphic gradient. Beneath the detachment faults, granites and gneiss have gently dipping foliations and have been extensively mylonitized, with late cataclasis reflecting the change from ductile to brittle deformation during decompression and cooling as the normal faulting in overlying layers exhumed the deeper rocks. Extensional metamorphic core complexes have now been recognised in all orogenic systems as late collapse structures formed during the waning stages of collisional orogens.

Laboratory experiments on analogue models have investigated extension of two-layer brittle / ductile systems. In these experiments, the brittle upper crust is represented by a sand layer, a Mohr-Coulomb material with a mean 30° friction angle; the behaviour of the ductile lower crust is simulated by silicone putty, a Newtonian fluid of 10^4 Pa.s viscosity at room temperature. Faults initiate as steeply dipping normal faults (60°) defining a few grabens and tilted blocks in the brittle layer. No fault offsets the initial brittle-ductile interface. This interface becomes a horizontal décollement at the onset of extension to accommodate the shear due to the relative displacement of the overlying brittle crust blocks.

Development of a detachment system
model after Brun *et al.* (1994) *Geology* **22**, 319-322



The steep normal faults progressively rotate to lower dips while continuing extension takes place in regularly spaced sites. This evolution leaves virtually undeformed areas between the extending sites, the result looking like crustal-scale pinch-and-swell boudinage. Heterogeneous deformation of the brittle layer is accommodated by pervasive flow of the ductile layer. This large-scale behaviour leads to a main, upward convex detachment fault that allows the ductile layer to rise in a dome-like structure. Of structural interest is that one limb of the dome is bounded by the detachment zone whereas the other limb results from block rotation forming a **roll-under** of the footwall.

Diapirs

Diapirism and **intrusion** are processes involved when a geological formation (the **source layer**) has come under sufficient stress (including gravity driven components) to flow, pierce and break through overlying strata of higher density and lower viscosity. Diapir structures may or may not be associated with regional deformation.

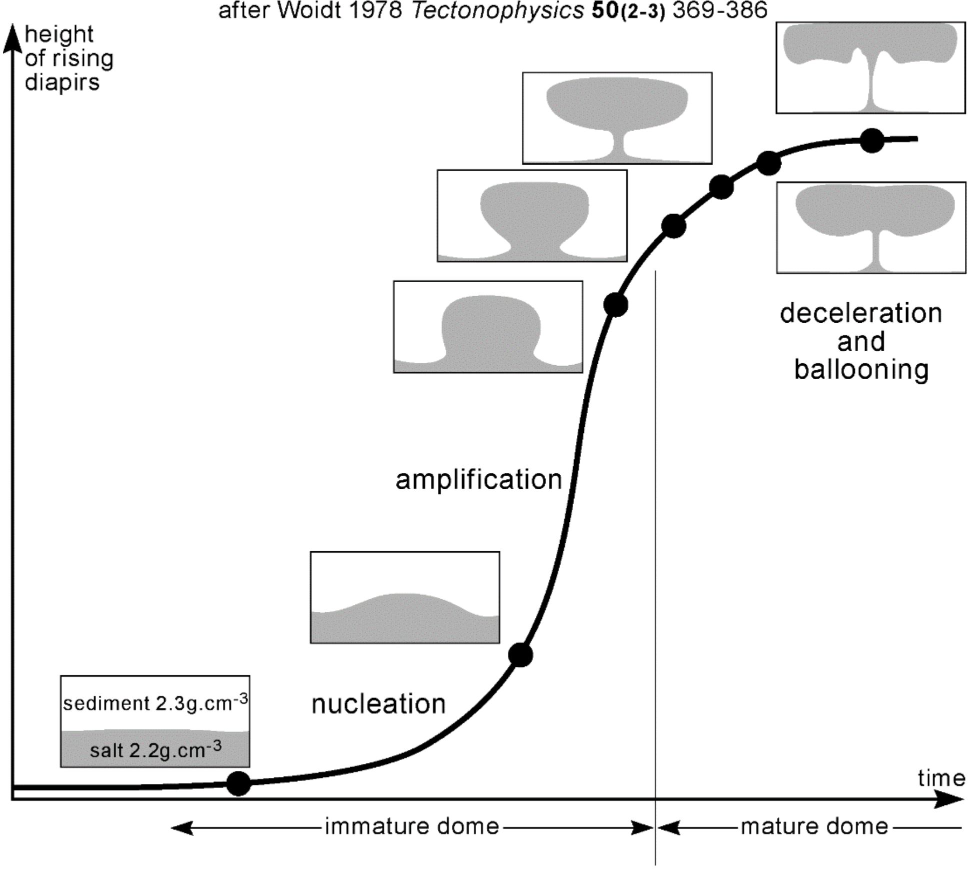
Models of diapirs consider two superposed layers of immiscible and viscous fluids of uniform density ρ_1 and ρ_2. Whether viscosities are equal or different, the setting is metastable if the overlying layer is denser than the lower one. Infinitesimally small perturbations on the horizontal interface are unstable and become amplified at a rate that depends on:

- the thickness, density and viscosity of the two layers
- the size of initial perturbations
- the time elapsed.

The **gravitational instability** of a heavy fluid overlying a lighter fluid is named Rayleigh-Taylor instability. It is independent of regional shortening or extension. The average spacing of such instabilities arises as a dominant wavelength phenomenon and is 2-3 times the depth of the mobilised, buoyant layer.

Geological diapiric structures are generally considered to be driven upward entirely by buoyant forces resulting from the density contrast between a low-density layer and the heavier rocks above it. As the low-density material rises, there is a complementary sinking of the overlying-higher-density material. For example, shale diapirs form where folding of unconsolidated sediments has taken place or where rapid sedimentation and compaction have generated high fluid pressures in unlithified shales and caused them to move upward. They form '**mud volcanoes**' where they reach the surface.
Fabric and strain patterns of diapirs have been determined experimentally and numerically. There are **internal structures** formed by flow in the intruding mass and **external structures** expressing the deformation of the country rock by the piercing diapir. The pattern of these secondary structures reflects the diapiric mechanism.

Evolution of a gravitational instability with viscosity contrast of 10
after Woidt 1978 *Tectonophysics* **50(2-3)** 369-386



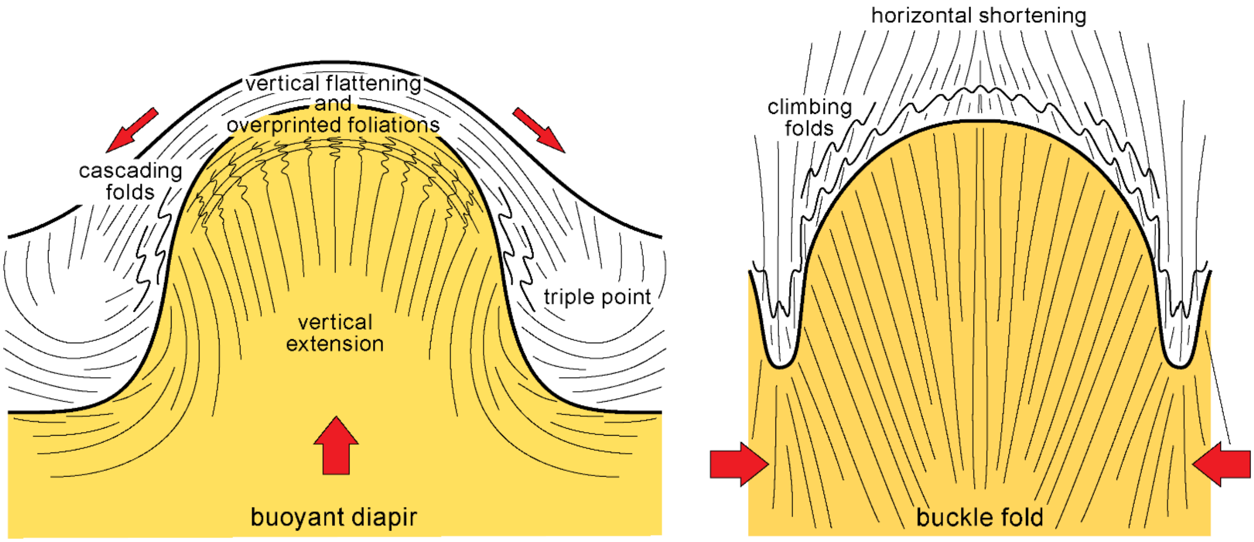
The essential differences with folds are:

- Layering in rock adjacent to the diapir is commonly dragged upward by rotation into the direction of diapiric movement. Folds around buoyant diapirs have a cascading vergence consistent with collapse of the hanging wall (core-side-up / mantle-side-down) and contrary to the classical S-Z shapes of second order folds climbing on a large-scale antiform. Low angle thrusts consistently indicate outward displacement of country rock. Marginal folds often have curved hinges caused by radial outward shortening as the diapir distends.
- A secondary foliation tangential to the diapir boundary dips outward less steeply than the core-/mantle contact. This foliation is axial planar to the cascade folds and bears a radial, downdip stretching lineation. In buckle folds secondary foliations would be steeper than the fold limbs.
- A very strong vertical flattening with a very strong horizontal extension dominates crestal regions of diapirs, whereas this region of anticlinal folds has horizontal shortening perpendicular to the axial plane foliation.
- Horizontal stretching in the crestal region of diapirs gives way at depth to very strong vertical stretching in the trunk region. As the diapirs grows the transition between the two regions moves

downward, with a resulting overprinting of horizontal fabrics on earlier vertical ones. Buckle folds do not exhibit this sequence of overprinting.

- In the cover sequence around magmatic diapirs, a very steep metamorphic gradient is preserved.
- Internal structures are characterised by doubly plunging, vertical sheath folds.

Comparative profiles through a diapir and a buckle fold with general patterns of secondary folds and foliation



model from Dixon J.M. in: Seyfert C.K. 1987
Encyclopedia of structural geology and plate tectonics, volume **10**, Van Nostrand Reinhold Company Inc. 398-412

Salt domes

Salt diapirs, conventionally called **salt domes**, exist in many places; well-known examples occur in the Persian Gulf and in Germany. Salt bodies usually contain varying amounts of other evaporites (especially anhydrite and its hydrated form, gypsum) and non-evaporite rocks that were originally interbedded with salt. Owing to similar shapes and apparent behaviour, salt bodies are often considered as analogue to igneous intrusions.

Salt properties

Salt and other evaporite material are sedimentary rocks that play special roles in geological deformation. Indeed, compared to other rocks:

- they have a low and almost depth independent density (2.15 to 2.2 g/cm^3, depending on the composition);
- they have a low equivalent viscosity (10^{18}-10^{21} Pa.s);
- they very easily become ductile and mobile;
- they behave almost like a Newtonian fluid over the geologic time scale and their movement is aided if a small amount of free water is present and/ or temperature increases;
- they are, as crystalline rocks, relatively incompressible.

Conditions for mobilisation

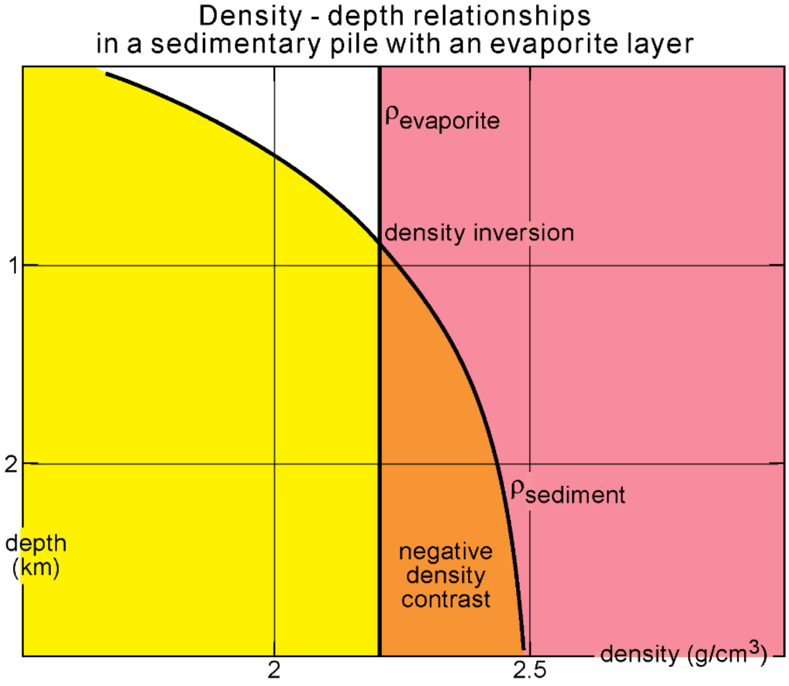
Driving conditions

For the salt movement to start, three conditions must be fulfilled:

- First, **density inversion** occurs below the **critical depth,** the depth below which evaporites become less dense than the overburden (a consequence of salt incompressibility vs. sediment compaction). Below this level, the system is gravitationally unstable. Evaporites become buoyant and will rise if the overburden strength is overcome.
- Second, sufficient load pressure is required to cause flow deformation of the salt. Empirically, it has been observed that salt layers about 300 m thick begin to react as a viscous fluid when their

cover reaches 1000 to 1500 m, the depth at which sediments become denser than salt. At depths of 1 - 3 km, salt will flow under low differential stresses due to its low strength. The rate of flow of salt on the geologic time scale averages 0.3 mm/a.

- Third, a disturbance is required to trigger the diapiric process. If the salt layer and the strata above are perfectly horizontal, regular and uniform, there is no gravitational instability. However, any lateral variation in the thickness of the overlying rock units will lead to differences in the pressure distribution on the salt layer. Salt will tend to flow towards those places where the pressure of the overburden is least. The flow of salt forced from high load areas is called **salt expulsion**. The initial impulse is often caused by a tectonic event, for example folding or faulting that impels displacement of one salt region relative to another, or by a thermal gradient, or by local erosion. This differential loading ultimately leads to the flow of all the salt to the surface. Buoyancy may still play an important role but is no longer considered the initiating factor in geological diapirism.



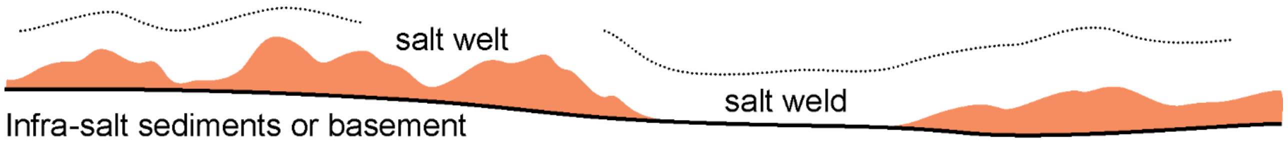
Restraining conditions

Two main factors limit the ability of salt to move into salt bodies:

- The strength of the overlying sediments;
- Boundary drag along the top and bottom surfaces of the salt layer.

Sedimentary rocks typically increase in both shear strength and frictional strength as depth of burial and confining pressure rise. Thick sedimentary roofs are therefore generally more difficult to deform than thin roofs. Roofs more than several hundred meters thick are unlikely to be deformed by salt of modest structural relief without assistance from either regional extension or shortening. Salt will flow only if gravity, thermal and tectonic driving forces exceed the resistance to flow.

The volumetric flux of laminar flow within a Newtonian salt layer has been estimated to be proportional to the third power of the layer thickness. This means, for instance, that halving the layer thickness slows flow by a factor of 8. Thus, the salt flows slower while salt expulsion reduces the thickness of the layer below until the viscous shear resistance may effectively immobilizes a remnant thin salt layers. Then a dome can no longer grow because the supply of nearby salt has become exhausted; exhaustion may cause the top and bottom contacts of the salt to merge, forming a **salt weld**.



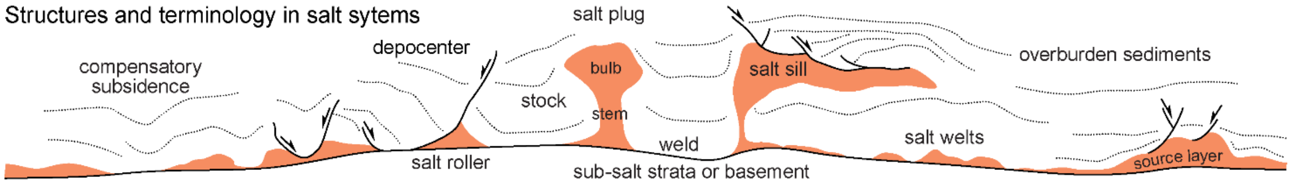
Structure of salt bodies

Numerical and analogue experiments have established that salt diapirs vary in shape as a function of depth. Position and shape both depend on how the brittle overburden deforms.

Shapes

A wide variety of three-dimensional forms reflects different stages in the upward migration of salt, commencing with simple broad anticlines or domes (**plugs** and **pillows**) and proceeding to **stacks, walls**, **columns**, **bulbs** and **mushroom** shaped forms. The raising part may become detached from the low-density source layer, in particular during regional shortening that pinches off the waist; the bottom part of the resulting **teardrop** diapir remains at depth to form a salt **pedestal**. Strata above and below the source salt layer may come in contact at a **weld**, where all the salt has flown away.

Salt domes are typically a few km in diameter and have very steeply dipping sides that may extend downward for several kilometres to form a relatively narrow **neck** that exhibits very tight folding with a strongly developed vertical linear elongation. Fully amplified salt domes extrude onto and may spread over the topographic surface to form salt **glaciers**.



Steep sides and flat tops, almost cylinder-like diapirs are shallow. Shallow to moderately deep diapirs tend to display rounded tops and less steeply inclined flanks. Deep salt diapirs have round, dome-like tops and relatively gently sloping sides.

External structure

The salt moving upward in a ductile manner pierces through the overlying sediments. Any rock occupying the space taken by the salt body must be removed or displaced by faulting and/or uplift of the overburden and/or erosion. In extensional systems, diapirs rise up fracture zones, taking advantage of the space created by thinning and separation of fault blocks.

Diapir roof

The strata above the salt body are affected by extensional tectonics reflected in arching and thinning of layers and in radial and concentric normal faulting. **Flaps** of roof rocks may be lifted, rotated and shouldered aside as the diapir amplifies.

Rim syncline

Boundary-drag along the margins of the diapir bends upward the sediments, which are truncated against the salt. Upward bending of the strata along the walls may be accompanied by reverse faults. The faulted and sheared sediments commonly form a clay-rich gouge against the salt: the **shale sheath**.

The sinking sediments over the area from which the salt has migrated often produce a **rim syncline** surrounding the diapir. In these synclines sedimentation compensates subsidence. Ultimately, a ring-shaped zone of compressional folding or faulting surrounds the essentially cylindrical neck of the salt intrusion.

The propagation direction of the salt is centripetal inward and upward and the converging motion generates peripheral tangential stresses. Thus, numerous folds whose axes trend radially and plunge outward from the dome in all directions originate at the base of the dome.

Internal structure

In general, salt domes do not move as a single mass but as a series of "**spines**" (i.e. smaller, possibly cylinder-shaped blocks) that move independently while intense deformation zones at their boundaries accommodate their relative movements. Deformation features of the salt include rotated boudins, stretched desiccation polygons and foliation formed through deep flowage of the salt, and folding and bedding-plane slip developed at shallow depths.

Caprock

In arid environments, salt domes that reach the surface flow out (as in the Persian Gulf). In moist environments, solution of the salt atop the rising diapir combines with residual accumulation, concentration, secondary alteration and precipitation of insoluble and little-soluble material such as anhydrite, gypsum, shales, carbonates and sulphurs to form an irregular complex of brecciated cavernous **caprocks**. The bottom boundary of the residual caprocks is a dissolution level that cuts through the head of the diapir. Caprocks are relatively impermeable and form a seal that blocks fluid (e.g. oil) migration. They occasionally extend along the sides of mature diapirs. Caprocks may contain entrained blocks of country rocks.

Trunk

The boundary with sediments is strongly sheared in the **salt sheath**. The internal structure of the salt is dominated at middle levels by very tight folds with near-vertical axes and axial planes, with foliation. Refolding or multiple refolding is common and is dictated by the direction and extent of flow. This may be connected in part with episodic upward movement of the salt, which is believed to rise in a complex fashion by the advance of local spines and lobes (multiple bulbs). A vertical flow lineation is common.

Once salt reaches the surface, it can continue to rise by passive diapirism, in which the diapir grows as sediments accumulate around it. A rapidly rising passive diapir may spread over the sediment surface to form an allochthonous salt sheet. A variety of salt-sheet lineages are possible, depending on the geometry of the feeder and the tectonic setting.

Salt-dome salt, like glacier ice, has undergone extensive flow and recrystallisation. Study of salt structures may therefore help to establish whether or not certain bodies of deformed silicate rocks (such as the migmatitic cores of gneiss domes) have also been emplaced and deformed by buoyant forces.

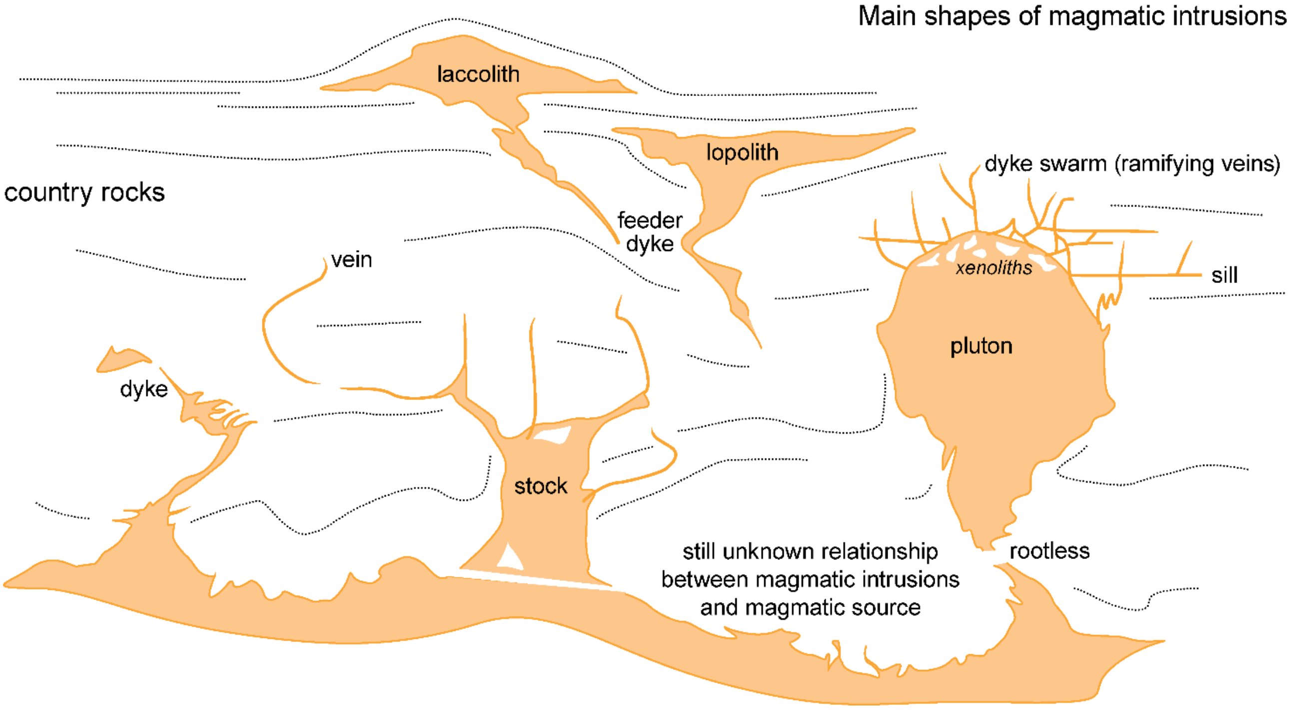
Mode of emplacement of plutonic bodies

Igneous diapirism is probably an important process in the vertical motion of magma through the lithosphere. Indeed, some plutons are similar to salt domes in shape and wall-rock deformation features. Magmas commonly have densities lower than those of the overlying rocks and consequently tend to ascend through passageways or zones of weakness. Most magma does not reach the Earth's surface but crystallises at depth to form **plutonic bodies** of igneous rocks. If plutons are diapirs (a concept still disputed) magma stops rising where surrounding rocks have lower density and/or at the temperature-equivalent depth where magma cools and solidifies.

Shape of magmatic bodies

The shape and orientation of many plutonic bodies is a direct result of the pre-existing structure of the **country rock** into which they have emplaced. The forces that give rise to deformation also control the way magma is emplaced and, to some extent determine the geometry of the resulting igneous body. Whereas plutonic bodies come in all sizes, and in a wide variety of forms, some forms are sufficiently recurrent to allow a classification on the basis of shape.

- **Dykes** and **sills** are tabular, approximately parallel-sided bodies that are thin in relation to their along-strike extent. Dykes look like walls that vertically or steeply cut across surrounding rock units. Many dykes intruded next to each other form a **swarm**. Sills are horizontal or low angled bodies concordant (i.e. sub-parallel) to the country-rock fabric or bedding-planes. They are usually thicker than dykes. Dykes and sills are commonly finer-grained or even glassy at their margins (**chilled margin**). The symmetrical character of chilled margins and contact metamorphism (top and down), and upward **apophyses** allow distinguishing sills from lava flows. Dykes and sills probably contain magma that has raised and quickly solidified in the relatively cold and brittle crust. They indicate that magma may rise through crustal fractures.
- **Ring dikes** are concentric and cylindrical, often with an outward dip. **Cone-sheets** are concentrically inclined, downward converging dykes.
- **Pluton** is the general name given to a relatively large body of intrusive igneous rock.
- **Laccoliths** are thick, lens-shaped sills with a flat floor and a convex upwards upper surface. Conversely, **lopoliths** are thick, large and lens shaped bodies with a flat roof and a convex downward lower boundary.
- Steep sided and discordant bodies are **plugs** (rather small and cylindrical, e.g. **necks**) **stocks** (larger than plugs) and **batholiths**.



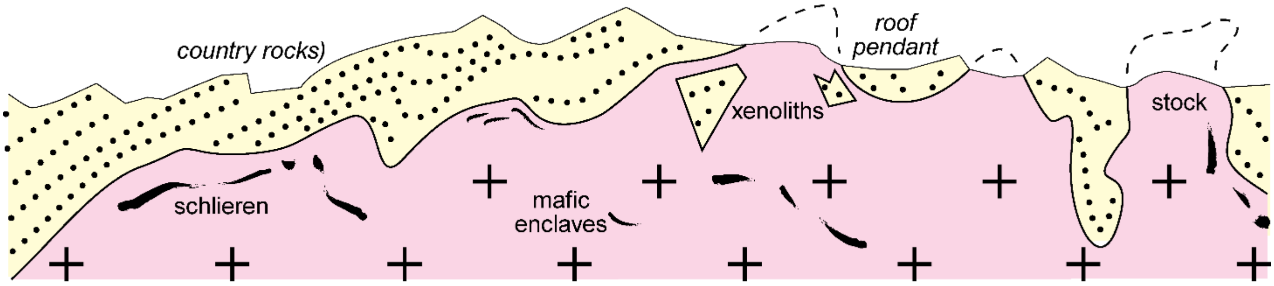
Attention: This classification may depend on the depth of erosion, the small size of plugs and stocks possibly representing the top of a larger batholith.

Batholith is a term that embraces very large units of plutonic rocks. They actually are composite masses. Diversity results either from magmatic differentiation of a single intrusion or from repeated multiple intrusions of many small and coalescing plutons. The mode of emplacement, for example as to whether they push aside the surrounding rocks, is still debated.

- In the roof of magmatic diapirs, extension joints are commonly filled with magmatic rock to form **dike swarms**. They reflect bending-related extension during doming of overlying strata.
- A **Caldera** is a large circular depression formed when the roof of a volcanically emptied magma chamber has collapsed,

Magmatic structures

It is thought that, during emplacement, magma is more likely a pasty mush than a free-flowing liquid. Crystals forming a solid are submitted to the bulk flow of the rising magma. Structures found within the igneous bodies are 1) the **igneous structures** formed by the movement of magma during emplacement, and 2) later structures due to deformation of the crystallised solid rock. Igneous structures include a **flow foliation** and a **flow lineation** essentially formed by the alignment, during the flow of the magma, of tabular and linear porphyroblasts crystallised before the final consolidation of the magma. A **flow banding** or **lamination,** which may include some **schlieren**, may be formed in viscous magmas. Stretched pieces of usually slightly more basic magma, yet containing the same minerals as the main rock, are also included in the banding.



Magmatic structures in a pluton with concordant and discordant contacts

These structures permit to infer the direction of flow of the fluid material within the chamber of the intrusion and, to some extent, the general shape of the intrusion because the foliation tends to be parallel to the intrusion / country rock interface. Many large intrusions exhibit a foliation which is parallel to their margin and which decreases in intensity towards the interior of the body. Studies of shapes of xenoliths show that this decrease in intensity is related to a regular pattern of deformation attributed to **ballooning** effect of later pulses of magma on an earlier, partly or wholly consolidated material. However, some of this deformation can be attributed to the deformation brought about by upward diapiric flow under gravitational pressure.

External structures express the deformation of the country rock by the intrusive action of the igneous body.

Intrusion processes

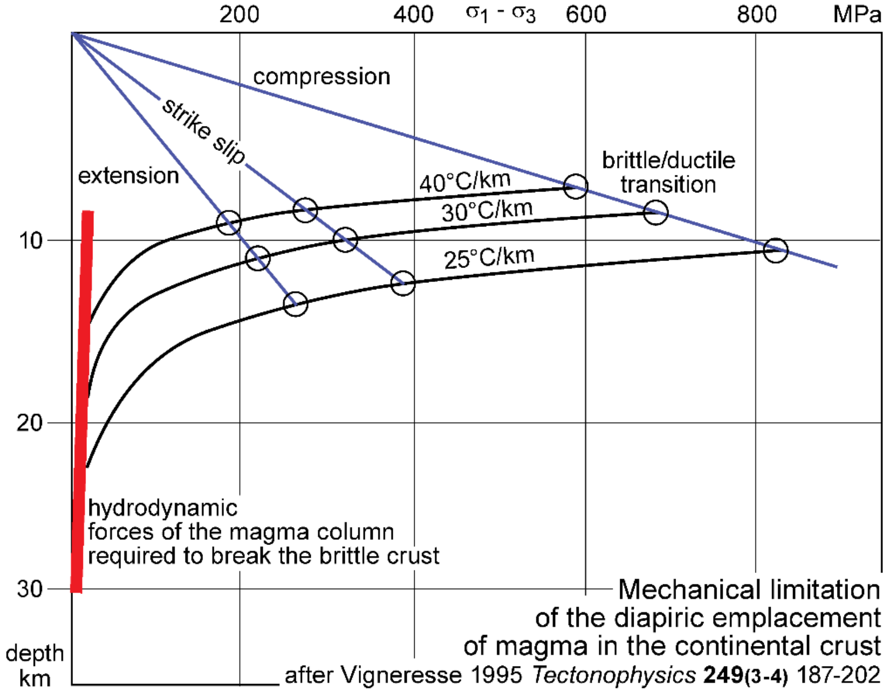
The general structural problem of **intrusion** (as a process) is how large magmatic bodies became emplaced within the crust and, in particular, how the space they occupy was created. This is the so-called **space problem**.

For some time some geologists proposed that the country rock had been somehow transformed (**granitized**) to produce the igneous magma. It is now acknowledged that large igneous bodies are magma that has ascented through parts of the crust (if not the lithosphere). There are two main views on the emplacement mechanisms:

- Plutons result from diking. Magma invades fractures and blocks of country rock detached from the roof and walls of the igneous body founder into the rising magma. These incorporated pieces of country rock (called **xenoliths**) may experience varying degrees of melting or chemical dissolution. This process is described as **stopping**. In this case, plutons are emplaced as **permitted intrusions**, and the structural relationships suggest a passive accommodation of the intrusive magma to the space left by the country rocks as it moves aside or subsides below the intrusive body. Some **ring-shaped** bodies were regarded as permitted by the subsidence of a central, cylindrical block into a rising, passively emplaced pluton.
- Plutons result from diapirism. Plutons are emplaced as **forceful intrusions** making space for themselves by actively pushing aside the surrounding country rock. The volume of a pluton is accommodated either by mechanical deformation in the pushed-aside country rocks or by

engulfment of country rocks, or by both sets of processes operating concurrently, country rocks being in part at least, digested or assimilated by the magma. Plutons are formed by magma spreading or swelling outwards (**balooning**) from a relatively narrow feeding neck at a particular level of the crust. However, some large magmatic plutons display fabric patterns similar to mantle gneiss domes, which is interpreted in terms of inflation of the plutons by successive magmatic pulses, each of which leading to stretching of the previously consolidated igneous material (balloon tectonics). The main structural features for diapirism are:

- Sub-circular shapes
- Parallel, marginal foliations in both the pluton and the host rocks
- Increase of the strain intensity from the core to the margins of the pluton
- Horizontal extension in the upper parts but steep stretching lineations and shear components along the margins
- Important deformation of the country rocks. Strain envelopes are consistent with tangential stretching, while less extreme manifestations include rim-concentric folds and radial fracture patterns.



Mechanical limitation of the diapiric emplacement of magma in the continental crust

after Vigneresse 1995 *Tectonophysics* **249**(3-4) 187-202

Models

The driving force of diapirism is the density inversion. The spontaneous rise of buoyant domes into a denser overburden in the Earth's gravity field has been examined theoretically and experimentally.

Analytical approach

The rock system is assumed to consist of several horizontal layers, each of which has uniform thickness, density and Newtonian viscosity. The layer interfaces separating the denser overlying layer from the lower-density substratum possesses infinitesimally small sinusoidal deflections of many wavelengths. Each perturbation is capable of growing in amplitude, due to the density inversion, and the mathematical treatment allows calculation of the relative rates of amplification of these wavelengths. The perturbations that grow at the maximum rate in time dominate the system. They develop into high amplitude ridges or domes and their wavelength are called the dominant wavelength of the system.

Early work by Rayleigh assumed no surface tension. He demonstrated that the exponential growth rate of the interface is proportional to $k^{1/2}$, a number given by:

$$k = 2\pi L / \lambda$$

where L is the length of the model and λ the wavelength of the interface instability.
Initial perturbations grow exponentially with time according to the relation:

$$A(t) = A_0 \exp(qt)$$

where A_0 is the initial amplitude, q is the growth rate and t is time.
Taylor confirmed this result, which is valid only for the nucleation stage of the instability (small amplitude with respect to wavelength). Later studies have included effects of surface tension, showing that the growth rate q is then given by:

$$q^2 = \left(\frac{\rho_2 - \rho_1}{\rho_2 + \rho_1}\right) gk - \left(\frac{s}{\rho_2 + \rho_1}\right) k^3$$

where ρ_1 and ρ_2 are the densities of the lighter and heavier fluids, respectively, g the gravitational acceleration and s the coefficient of surface tension. According to this equation, the configuration is stable if $\rho_2 < \rho_1$ and unstable if $\rho_2 > \rho_1$ for all wave numbers $0 < k < k_c$ where k_c is the cutoff wave number given by:

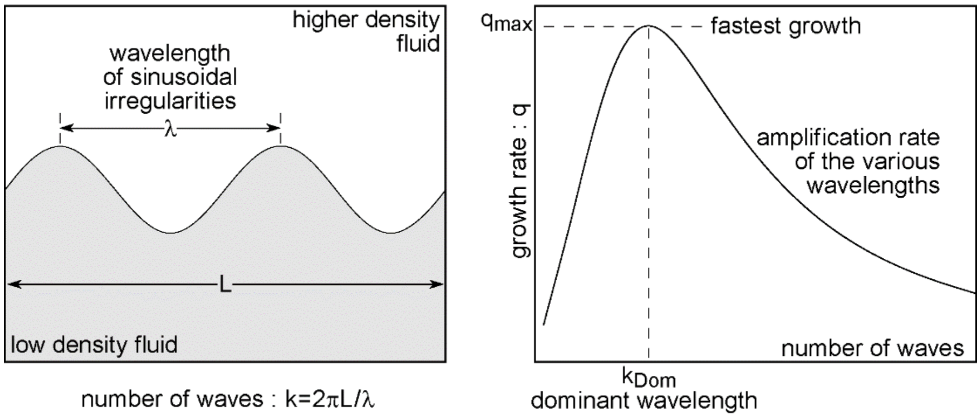
$$k_c = \sqrt{g(\rho_2 - \rho_1)/s}$$

The effect of surface tension is to stabilize all wave numbers larger than k_c. In addition, there is wave number k_{dom} for which the instability grows fastest:

$$k_{dom} = k_c / \sqrt{3}$$

Mathematical model for diapirism

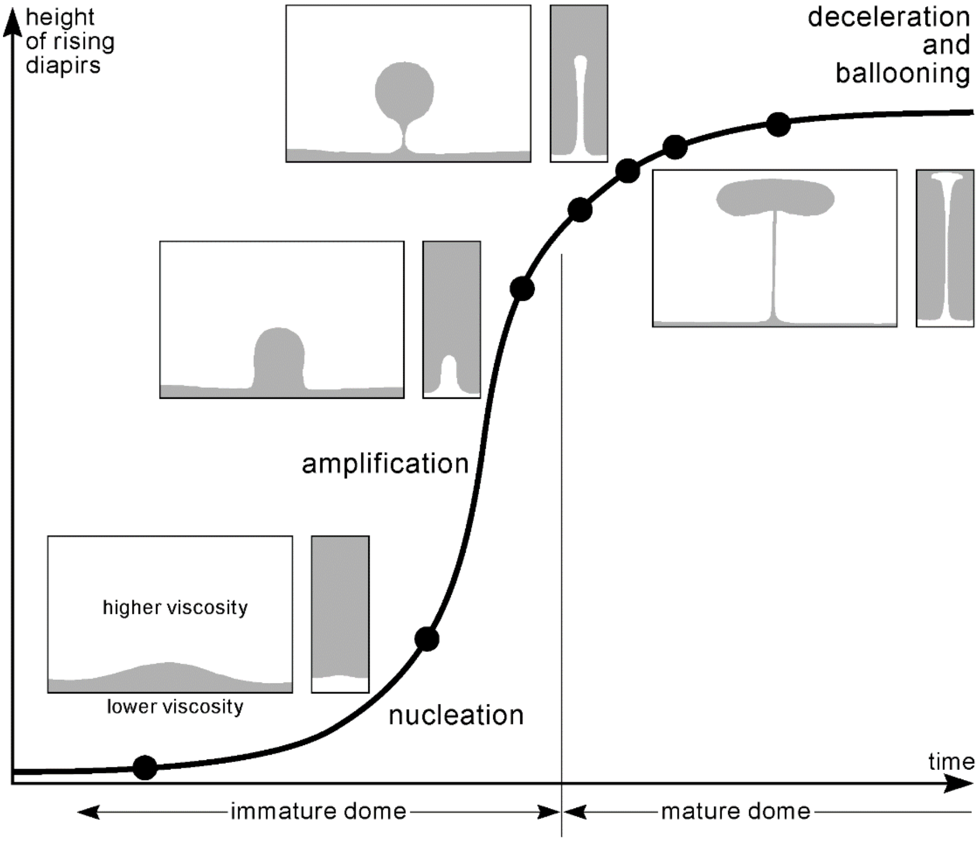
Absolute values of the growth rate q_{max} and the dominant wavelength k_{Dom} depend on the rheologie of the fluids, surface stresses and boundary conditions
e.g. Turcotte & Schubert, 1982 *Geodynamics: Application of continuum physics to geological problems* John Wiley, New York, 450p.



Note that in all of these equations the wavelength numbers are independent of the viscosity. The effects of viscosity and other parameters or physical variables (e.g. density gradients are similar to those of surface tension, producing dominant wavelengths and stabilizing larger wavelength numbers. The dominant wavelength and the rate of amplification of the dominant perturbations are functions of the thickness, viscosity and densities of the material layers in the system. In relative terms, for a fixed set of buoyancy properties, the following effects are predicted:

- increasing the relative thickness of the overburden yields a larger dominant wavelength and a faster amplification rate

- increasing the relative viscosity of the overburden yields a larger wavelength and a slower rate of amplification
- Increasing the density contrast has an almost negligible effect on the wavelength but dramatically increases the rate of amplification.



Effect of viscosity on the development of a gravitational instability
after Woidt 1978 *Tectonophysics* **50(2-3)** 369-386

Most of the theoretical work on pluton emplacement approximated the plutons as Newtonian spheres and tried to calculate their ascent time depending on the rheology of overburden. This is the Stokes flow problem. If the surrounding material is Newtonian, the vertical velocity of the sphere is given by:

$$v_z = \frac{\left(\rho_{host} - \rho_{sphere}\right) g.r^2}{3\mu_{host}} \left(\frac{\mu_{host} + \mu_{sphere}}{\mu_{host} + \left(3/2\right)\mu_{sphere}} \right)$$

where ρ are densities, μ viscositie, g the gravitation acceleration and r the radius of the sphere. When $\mu_{sphere} <<< \mu_{host}$ the equation is simplified to the Stokes formula:

$$v_z = \frac{\left(\rho_{host} - \rho_{sphere}\right) g.r^2}{3\mu_{host}}$$

This equation used with standard physical properties of rocks shows that diapiric plutons are emplaced between a few 10^4 to 10^5 years.

Hydraulic head

Gravitational loading is a combination of the weight of rocks overlying the salt and the gravitational body forces within the salt. It is incorrect to say that salt like fluids flow in response to pressure gradients. A classic example is that water in a pond does not flow downward although the pressure increases downward. The mobilizing effects of gravitational loading are thus approached through the concept of **hydraulic head gradient**. All fluids flow from zones of high head to low head; if the hydraulic head is everywhere the same, the fluid remains at rest.

The total hydraulic head has two components: (1) elevation head and (2) pressure head.

- Elevation head is the elevation of a fluid particle above an arbitrary horizontal datum.
- Pressure head is the height of a fluid column that could be supported by the pressure exerted by the overlying rock.

Mathematically,

$$h_h = z + \frac{P}{\rho_S g}$$

where h_h is the total head, z is the elevation above a horizontal datum, P is the lithostatic pressure exerted by the overburden, ρ_S is the salt density, and g is the gravity acceleration. Substituting the lithostatic pressure with $\rho_R gD$, this equation simplifies to:

$$h_h = z + \frac{\rho_R}{\rho_S} D$$

where ρ_R is the average density of overlying rocks and D (depth) is the thickness of the overburden. Assuming that salt behaves as a fluid in response to gravitational loading over geologic times, flow directions can easily be predicted by estimating the head gradient for simple geologic situations.

Experimental approach

The evolution of buoyant domes has been simulated in details. A scale model is constructed of viscous rock analogues (e.g. silicone putty) and the model may be submitted to increased body forces in a centrifuge. Such models allow following the evolution of strain patterns in complex systems. For simple systems, the mathematical solution is now sophisticated to a point that it provides better answer than analogue modelling, in particular for cases such as diapirism in the mantle, which leads to hot spots, because the thermal effects are difficult to account for in analogue systems.

Conclusions

Large scale closed structures termed domes and basins have various origins. Plutonism is an obvious mechanism. Metamorphic core complexes are rocks of middle crustal levels that were subsequently uplifted and exposed by a process dominated by regional extension. Conversely, mantled gneiss domes are cored by rocks of middle crustal levels that were brought to shallower levels by buoyant diapirism and/or polyphase folding and erosion without the necessity of horizontal extension.

For all of these structures, gravitational instability and diapiric rise seems to play a significant role. However, buoyancy is no longer considered to be important factor in initiating diapirism. Differential loading is the dominant force overcoming the strength of the overburden and boundary friction, hence allowing salt or magma to flow. Otherwise, salt can remain static in the subsurface for tens or even hundreds of millions of years, subject only to groundwater dissolution, diagenesis, and metamorphism.

It is not established, however, whether dome and basin structure is necessarily associated with diapiric behaviour or whether the domes and basins represent embryonic folds that are just starting to be amplified in a more-or-less undeformed sheet.

Movies

http://www.beg.utexas.edu/indassoc/agl/agl_if.html

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