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Cross-references

Deep Seismic Reflection and Refraction Profiling
Energy Budget of the Earth
Heat Flow Measurements, Continental
Heat Flow, Continental
Heat Flow, Seafloor: Methods and Observations
Lithosphere, Continental
Lithosphere, Mechanical Properties
Lithosphere, Oceanic: Thermal Structure
Mantle Convection
Radiogenic Heat Production of Rocks
Sedimentary Basins
Seismic Velocity-Temperature Relationships

LITHOSPHERE, MECHANICAL PROPERTIES

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Synonyms

Geosphere

Definition

Lithosphere (*mechanical*). This is the rigid ("litho" = stone) outer layer of the Earth that remains mechanically strong over geological time spans. *Mechanical lithosphere* includes rigid layers of crust and outermost mantle capable to maintain high differential tectonic stresses, from 10 MPa to 1 GPa. *Mechanical lithosphere* is 1.5–2 times thinner than "seismic," "thermal," or "chemical" lithosphere.

Mechanical properties of the lithosphere. This term refers to the *integrated strength* of lithospheric plates, their *rheological structure* and parameters, and mechanical behavior in response to various tectonic loads.

Introduction

Mechanical properties of the lithosphere are of primary importance for local and global geodynamics. In particular, compared to the convective mantle, high long-term mechanical strength makes the lithosphere a unique stress/strain guiding and accumulating envelope with lasting mechanical memory. High strength prohibits internal *heat advection*, so *thermal conduction* is main heat transfer mechanism in the lithosphere, in contrast to the convective mantle. High strength also stabilizes vertical lithological structure of lithosphere making it a *stagnant layer*. In contrast to viscous mantle, long-term *rheology* of the lithosphere is strongly influenced not only by its *ductile* but equally *elastic* and *brittle* properties. It is probably the nonviscous properties of the lithosphere that shape it in the characteristic *plate tectonics* patterns.

The term *lithosphere* has been introduced in the second half of the nineteenth century, while the notion of *mechanical lithosphere* appeared in early twentieth century, in conjunction with that of *seismic lithosphere* (see *Earth's Structure, Global*), after formulation of the *continental drift theory* by Wegener and first interpretations of *regional isostasy* by J. Barrel and Vening-Meinesz (Watts, 2001; see entry *Isostasy*). The fact that the lithosphere has finite measurable strength has been demonstrated from observations and models of *regional isostatic compensation* of large topographic loads. Before that, the lithosphere was considered either as a very strong solid layer (*Pratt's model*) or, in turn, a weak fractured layer (*Airy's model*). *Postglacial rebound studies* of early twentieth century have contributed to the definition of the *mechanical lithosphere* as the uppermost layer of the solid earth characterized by slow *viscoelastic relaxation*, in contrast to the underlying, relatively low-viscosity asthenosphere. The long-term mechanical base of the lithosphere, h_m, is limited by the depth to isotherm 500–600°C in oceans and 700–800°C in continents, compared to almost twice as deep 1,330°C isotherm delimiting the thermal lithosphere (see entries *Lithosphere, Oceanic: Thermal Structure*; *Lithosphere, Continental: Thermal Structure*). As suggested on the basis of recent mantle–lithosphere interaction models (e.g., Schmeling et al., 2008), it is the elastic and plastic properties of the lithosphere that essentially determine the geometry of lithospheric plates and the mechanisms of formation of constructive, destructive, and transform plate boundaries at global scale. At smaller scale, *mechanical properties* of the lithosphere control formation and evolution of major geological structures such as rifts, passive margins, foreland basins, mountain ranges, plateau or strike-slip faults. They also control short-term processes such as seismicity (Watts and Burov, 2003).

Mechanical properties at different timescales

Mechanical properties of the lithosphere are timescale dependent (e.g., Watts, 2001). At seismic timescales (<0.5 h), the lithosphere behaves as an elastic or brittle-elastic layer of a thickness largely exceeding h_m. At *post-seismic timescales* (from days to years), the uppermost crust relaxes a part of seismically induced stress evoking some rather poorly understood short-timescale viscoelastic properties. At *postglacial rebound* timescales (~several 1,000 year), lithosphere remains largely elastic, with thickness of the elastic domain almost as important as its *seismic thickness*. At geodynamic timescales

(>1 Myr), the thickness of the *elastic core* is reduced to values smaller than h_m, and the stresses are slowly relaxed at timescales on the order of several Myr. The long-term properties of the lithosphere were first assessed from geodynamic-scale observations, first of all, *regional isostasy*. According to these data, the lithosphere exhibits a large spectrum of long-term behavior from quasi-elastic to brittle and viscous. Observation of crustal and lithosphere scale faults and *distributed seismicity* shows that some domains within the lithosphere deform in brittle-plastic regime over long time spans, while relaxation of deformation below, for example, oceanic volcanic islands, points to long-term viscoelastic strength. However, the long-term properties of the lithosphere are more generally studied indirectly using extrapolations from *rock mechanics data*.

Sources of information on the mechanical properties of the lithosphere

Flexural studies and *experimental rock mechanics data* are major sources of quantitative information on long-term behavior and strength of the lithosphere. In structured *viscous-elastic-plastic* media, all rheological properties are interrelated, and various time- and scale-dependent factors may cause variations in the effective elastic, brittle, and ductile deformation. Consequently, observations of long-term deformation are of primary importance for assessment of the effective mechanical properties of the lithosphere. These and other sources of information are summarized below:

1. Observations of long-term ($t > 1{,}000$ year) lithosphere response to tectonic loading, in the order of importance:
   - Observations of *regional isostatic compensation*: *gravity-flexural studies* providing estimates for the *equivalent elastic thickness* of the lithosphere, T_e
   - Vertical motions due to *postglacial rebound*, *lake, and volcanic island* loading
   - Observations of *lithosphere folding* (*folding wavelength* is a function of plate strength)
   - Field observations of ductile and brittle deformation in the outcrops
   - Interpretations of *deformational microstructures* and *paleo-stresses*
   - *Seismic tomography* (e.g., evidence for more or less strong slabs at depth)
   - *Borehole* stress measurements
2. Intermediate timescale observations ($t > 3$–10 year):
   - *Slow-rate rock mechanics experiments* (strain rates 10^{-9} to 10^{-4} s^{-1}, for ductile properties)
   - Geodetic (GPS-INSAR) data over >5 year time spans (strains, viscoelastic properties)
   - *Inter-seismic deformation*; *slow earthquake* data
3. Observations of deformation in response to short-term loading ($t < 1$ year):
   - Short-term *rock mechanics experiments* (elastic and brittle properties)
   - Distribution of intraplate seismicity (brittle properties)
   - Tidal deformation (viscoelastic properties)
   - Post-seismic relaxation (viscoelastic properties)
   - Geodetic (GPS-INSAR) data (strains, viscoelastic properties)
   - Attenuation of S waves (proxy to low-viscosity zones)
   - Magnetotelluric sounding (reduced electrical resistivity is proxy to low-viscosity zones)
4. Physical considerations and self-consistent thermo-mechanical models:
   - Estimates of the minimal *integrated strength* of the lithosphere required for lifetime stability of geological structures, *subduction* or transmission of tectonic stresses, and forces over large spatial scales, including horizontal pressure gradients caused by lateral variations in lithospheric density structure and topography (*gravity potential energy* theory). For example, lithosphere must be strong enough to transmit *ridge push* and *slab pull* forces on the order of 10^{11}–10^{13} N per unit length.
   - Lithosphere scale *numerical thermo-mechanical models* of tectonic processes integrating multidisciplinary data, which allows for testing the validity of data and hypotheses on lithosphere rheology.

Observations of long-term deformation provide key parameters such as the *integrated strength* of the lithosphere. These parameters are needed to constrain *rock mechanics data* obtained at laboratory conditions, because they are too far from geological conditions: short timescales ($t < 5$ years), small spatial scales ($l \sim 0.1$ m), high strain rates $\left(\dot{\varepsilon} > 10^{-9}\ \mathrm{s}^{-1}\right)$, small strains ($\varepsilon < 10$), high temperatures, simple deformation, largely mono-phase samples. *Rock mechanics data* allow only for assessment of general form of rheology laws, their sensitivities, and relative strengths of different kinds of rocks. Their extrapolation to geodynamic scales ($t > 10^6$ years, $l \sim 1{,}000$ km, $\dot{\varepsilon} < 10^{-14}\ \mathrm{s}^{-1}$, $\varepsilon > 100$, cold temperatures, aggregate rocks) needs re-parameterization and validation using real-scale observations and models.

Interpretation of short-term data (seismic, satellite geodesy, INSAR) is not straightforward. In particular, interpretation of *intraplate seismicity* (see entry *Seismicity, Intraplate*) and *post-seismic relaxation* data is questioned due to the lack of evidence that mechanisms of this short-term deformation can be physically linked to those of long-term deformation.

Observations of flexural behavior and effective long-term strength of the lithosphere

Observations of *regional isostatic compensation* (e.g., Watts, 2001; see entry *Isostasy*) have shown that the lithosphere has substantial long-term *elastic rigidity* that allows for transferring and maintaining intraplate stress levels (10 MPa–1 GPa) over geodynamic time spans (>several Myrs). Studies of gravity anomalies observed over mountain ranges and subduction zones have demonstrated that lithospheric plates bend like *thin elastic plates* of finite stiffness in response to tectonic, topography, and sedimentary loads. With improvement of geophysical measurement techniques during the second half of twentieth century, multiple studies (specifically, forward *flexural models*) have produced robust estimates of *flexural rigidity*, *D*, and *equivalent elastic thickness*, T_e, of lithospheric plates. These data arguably present a major source of information on the long-term *mechanical properties* of lithosphere. *Flexural studies* reveal strong variations of lithosphere strength, from near zero at ocean ridges to 100–150 km thick quasi-elastic cores detected within old and cold cratons (Burov and Diament, 1995; Kirby and Swain, 2009). In *flexural models* of *regional isostatic compensation* (Watts, 2001), *D* is varied until the model-predicted basement or *Moho topography* provides optimal fit to observations. *Gravity data* (see entry *Gravity Anomalies, Interpretation*) are used when basement topography is hidden (e.g., by sediments) or not representative of flexure (e.g., modified by erosion). The most robust *gravity models* are *forward models*. *Inverse models* are widespread due to the ability to cover large zones, but their results are more sensible to errors and should be cross-checked with *forward models*. For example, if not properly formulated, *gravity admittance* techniques may generate spurious results in continents (e.g., Kirby and Swain, 2009).

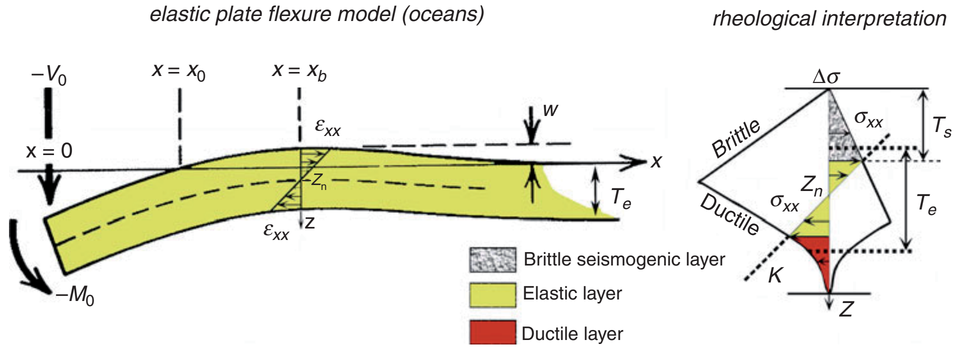
D provides a direct measure for the *integrated long-term strength* of the lithosphere and is linked to the *equivalent elastic thickness* of the lithosphere, $T_e : D = E\,{T_e}^3(12(1-\nu^2))^{-1}$, where *E* and *ν* are Young's modulus and Poisson's ratio, respectively. *Plate bending or flexure* is characterized by its vertical deflection, $w(x)$ and local radius of curvature, $R_x(x)$ or curvature, $K(x) = -R_x^{-1} = \partial^2 w/\partial x^2$ (Figure 1). The flexural equation, when expressed using *bending moment* $M_x(x)$ is rheology independent and is valid for all, elastic and inelastic plates:

$$\frac{\partial^2}{\partial x^2}\overbrace{\left(\underbrace{\frac{E\,T_e^3}{12(1-\nu^2)}}_{D(x)}\underbrace{\frac{\partial^2 w(x)}{\partial x^2}}_{K(x)}\right)}^{M_x(x)} + \frac{\partial}{\partial x}\left(F_x\frac{\partial w(x)}{\partial x}\right) + \Delta\rho g w(x) = \rho_c g h(x) + p(x) \quad (1)$$

where F_x is horizontal *fiber force*, $\Delta\rho$ is the density contrast between surface and subsurface material (i.e., between mantle density ρ_m and topography/sediment/water density ρ_c), $h(x)$ is topography elevation, and $p(x)$ is additional surface or subsurface load. The elasticity is used just as simplest rheological interpretation of bending strength: elastic bending stress is linear function of curvature and depth: $\sigma_{xx}(x,z) \approx (0.5T_e - z)EK(1-\nu^2)^{-1}$. T_e is thus effective parameter and should not be automatically related to any real layer within the lithosphere. For inelastic plates, T_e and *D* have a sense of "condensed" plate strength and are direct proxies for the long-term *integrated strength*, *B*, of the lithosphere (Watts, 2001). For example, for a single-layer oceanic plate ($T_e = T_{e\_ocean}$, Figure 1):

$$B = \int_0^{h_m} \sigma^f(x,z,t,\dot{\varepsilon})dz$$
$$\text{while } T_{e\_ocean} = \left(12\left(\frac{\partial \sigma^f_{xx}}{\partial y}\right)^{-1}\overbrace{\left(\int_0^{h_m}\sigma^f_{xx}(z-Z_n)dz\right)}^{M_x(x)}\right)^{\frac{1}{3}};$$
$$T_{e\_ocean} < h_m \quad (2)$$

where ${\sigma_{xx}}^f$ is brittle-elastic-ductile bending stress (Burov and Diament, 1995); T_e is usually much smaller than



Lithosphere, Mechanical Properties, Figure 1 Classical flexural model of oceanic lithosphere (*left*). *Right*: brittle-elastic-ductile *yield stress envelope* (YSE) and interpretation of the *equivalent elastic thickness* T_e of the lithosphere. $\varepsilon_{xx}(z)$ is *flexural* strain, $\sigma_{xx}(z)$ is *flexural stress*, $K(x)$ is local plate curvature, $\Delta\sigma$ is *differential stress*, $Z_n(x)$ is *neutral plain*, and T_s is brittle seismogenic layer. $w(x)$ is vertical plate deflection. V_0 and M_0 are boundary cutting force and *flexural moment*, respectively.

h_m. M_x and, hence, D can be obtained from depth integration of σ_{xx}^{f}. D and T_e may spatially vary due to their dependence on local bending that leads to localized plate weakening (called *plastic or ductile hinging*) in the areas of utmost flexure, for example, near *subduction zones* or below mountains and islands.

Rheological properties of lithosphere according to rock mechanics data

The long-term mechanical behavior of rocks is represented by *extended Maxwell solid*, in which total strain increment equals a sum of elastic, viscous (ductile), and plastic (brittle) increments while the elastic, viscous, and plastic stresses are mutually equal. The weakest rheological term thus defines the effective behavior of this solid. Goetze and Evans (1979) used *Maxwell solid* and rock mechanics data to introduce the *yield stress envelope* (YSE) of the lithosphere (Figures 1 and 2). This approach consists in predicting, for a representative *background strain* rate and depth–pressure–temperature profile, the maximal *yield strength* $\Delta\sigma_{\max}(z)$ as function of depth, z. If elastic *differential stress* estimate $\Delta\sigma^{e}(z) < \Delta\sigma_{\max}(z)$, the deformation is elastic and *differential stress* $\Delta\sigma(z) = \Delta\sigma^{e}(z)$. If $\Delta\sigma^{e}(z) \geq \Delta\sigma_{\max}(z)$, then $\Delta\sigma(z) = \Delta\sigma_{\max}(z)$ and the deformation is brittle or ductile depending on z.

Elastic properties

The *elastic behavior* is described by linear Hooke's law:

$$\sigma_{ij} = \lambda\varepsilon_{ii}\delta_{ij} + 2G\varepsilon_{ij} \tag{3}$$

where λ and G are Lame's constants. Repeating indexes mean summation, δ is Kronecker's operator. For most rocks $\lambda \approx G = 30$ GPa (Turcotte and Schubert, 2002).

Brittle-plastic properties

Brittle resistance, τ, is a linear function of normal stress, σ_n and, that is, of pressure (Byerlee, 1978):

$$\tau = 0.85\sigma_n, \quad \sigma_n \leq 200 \text{ MPa} \tag{4}$$

$$\tau = 50 \text{ MPa} + 0.6\sigma_n, \quad 1{,}700 \text{ MPa} > \sigma_n > 200 \text{ MPa}$$

Byerlee's law is equivalent of *Mohr-Coulomb* (*Coulomb-Navier*) *plastic failure criterion* (e.g., Burov, 2007):

$$\tau = C_0 + \tan(\phi)\sigma_n \tag{5}$$

where C_0 is *cohesive strength* (<20 MPa) and ϕ is the *internal friction angle* (30–33°).

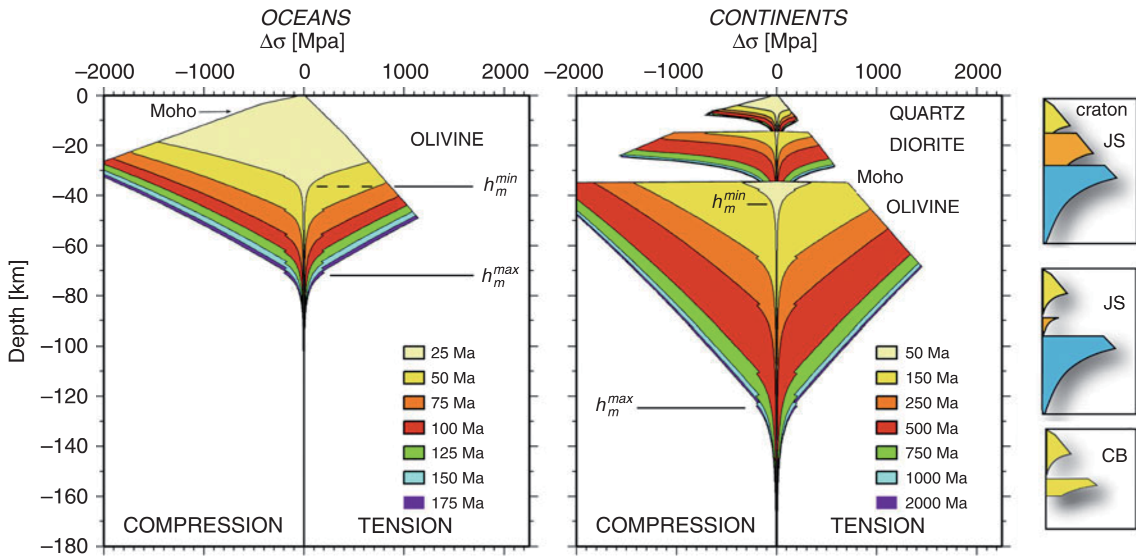
The Griffith criterion extends (5) to pure tensile domain:

$$\tau^2 = 4T_0^2 + 4T_0\sigma_n \tag{6}$$

where $T_0 = C_0/2$ is tension cut-off.

Viscous-ductile properties

Effective viscosity of the lithosphere varies from 10^{19} Pa s at the lithosphere–asthenosphere boundary to $>10^{26}$ Pa s in cold mantle near Moho depth. Several ductile mechanisms such as *diffusion creep*, *grain boundary sliding* (*GBS*), *pressure solution*, and *cataclastic flow* may play



Lithosphere, Mechanical Properties, Figure 2 Commonly inferred (brittle(Byerlee)-elastic-ductile) rheological *yield stress envelopes* (YSEs) as function of *thermotectonic age* for oceans and continents. In mantle, maximal strength can be limited by *Peierls* or *GBS* law instead of *Byerlee's law*. For continents, variations in crustal composition and fluid content result in various Jelly Sandwich, *Jelly Sandwich*, and more rare *Crème Brûlée*. After Burov (2007).

important role at appropriate conditions, but the leading part belongs to the *dislocation creep* (Kohlstedt et al., 1995):

$$\dot{\varepsilon}^d = Af_w \Delta\sigma^n \exp\left(-H(RT)^{-1}\right) \quad \text{for } \Delta\sigma < 200 \text{ MPa (Power law)}$$
$$\dot{\varepsilon}^d = Af_w \exp\left(-H\left(1 - \Delta\sigma/\sigma_p\right)^2 \Big/ RT\right) \quad \text{for } \Delta\sigma > 200 \text{ MPa (Harper} - \text{Dorn law)} \tag{7}$$

where $\dot{\varepsilon}^d$ is shear strain rate, A is material constant, n is power law constant, f_w is water fugacity factor, $\Delta\sigma$ is differential stress, R is universal gas constant, $H = Q + PV$ is creep activation enthalpy, Q is activation energy (100–600 kJ/mol), P is pressure, V is activation volume, T is temperature in K. σ_p is Peierls-like stress ($\sigma_p \sim$ several GPa). For tectonically relevant $\Delta\sigma/\sigma_p$ ratios (<0.1), Harper-Dorn's flow refers to *Peierls plasticity* that limits rock strength in high stress regime. *Dislocation creep* is strongly nonlinear non-Newtonian viscous flow ($n \sim 2$–4).

The second most important deformation mechanism, low-stress *diffusion creep* (Nabarro-Herring, Coble), results from directional diffusivity of rocks under applied stress (Kohlstedt et al., 1995):

$$\dot{\varepsilon}^d = Aa^{-m} f_w \Delta\sigma^n \exp\left(-H(RT)^{-1}\right) \tag{8}$$

where a is grain size and m is diffusion constant. For olivine, $m \sim 3$ and $n \sim 1$ and the constitutive law is linear Newtonian.

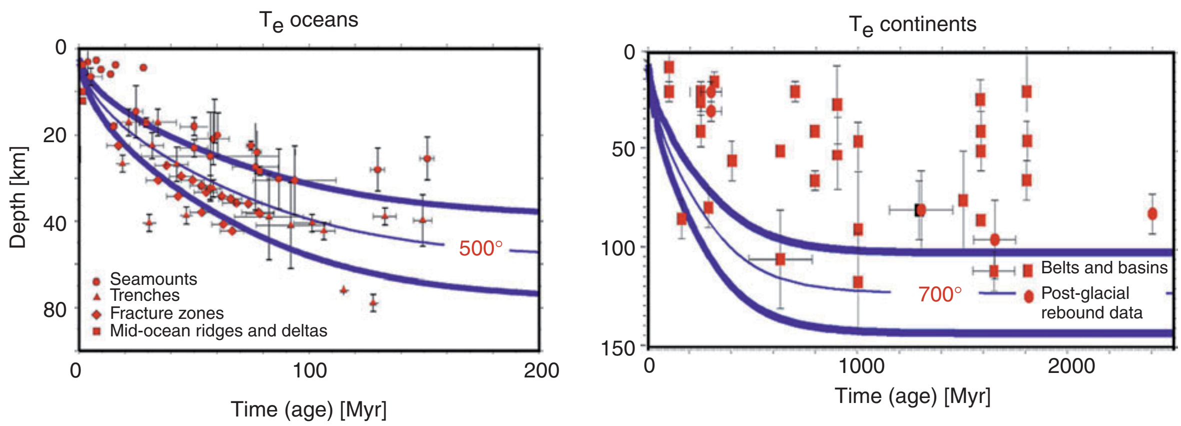
The *Peierls* super-plasticity is likely to replace *Byerlee's law* in the mantle below 30–40 km depth (Kameyama et al., 1999), that is, at high confining pressures (>1 GPa). Since *Peierls* plasticity includes strong water-weakening ($H = Q + p(V - \beta\Delta V_w)$), where ΔV_w is molar volume change due to incorporation of hydroxyl ions in the main rock and β is experimental parameter, it may play an important role in localization of deformation in subduction zones. This role may be shared with *GBS* creep, which might be responsible for aseismic localization of deformation in the lithosphere mantle.

Mechanical properties of oceanic lithosphere versus continental lithosphere

Due to its temperature dependence, ductile strength limits the depth to the base of mechanical lithosphere, h_m, to that of the isotherm 500–600°C in oceans and 700–800°C in continents (in continents, higher pressure for given temperature increases yield strength via the H term in Equation 7).

Oceanic YSEs ($0 < T_e < 50$–70 km) derived from (3) to (8) predict brittle seismogenic behavior in the upper parts of lithosphere, where depth of *brittle-ductile transition*, BDT, varies from few km near ridges to 40 km near subduction zones (Figures 2 and 3). The competent domains below the *elastic core* are dominated by aseismic ductile creep, where important ductile strength is preserved down to the depths of 80–100 km near subduction zones.

Continental YSEs (0–$5 < T_e < 150$ km) reflect strong *rheological stratification* between the upper, intermediate, lower crust and mantle lithosphere (Figures 2 and 3). There might be several BDT depths, typically at 15–25 and 30–45 km (Watts and Burov, 2003; Burov, 2007, 2010). The depth to the mechanical bottom of strong ductile lithosphere may vary from 30 to 200 km.



Lithosphere, Mechanical Properties, Figure 3 Observed T_e distribution in oceans and continents as function of *thermotectonic age* of the lithosphere, compared to depth of geotherms of 500 and 700°C defining, respectively, the mechanical base of oceanic and continental lithosphere (Burov and Diament, 1995; Watts, 2001; see also *Isostasy*). Assumed thermal thickness of continental lithosphere is 250 km. Depending on assumed thermal thickness and boundary conditions, predicted depth to 500–700°C may shift up or down by up to 25%. In continents, crust-mantle decoupling leads to ~50% T_e reduction compared to the depth to its mechanical base at 700°C. This explains occasional correlation of T_e with 400°C depth.

T_e estimates (Figure 3) reveal important difference between the mechanical properties of oceanic and continental lithosphere. In oceans, T_e (<70 km) follows the geotherm of $500 \pm 100°C$ and correlates with age. It fits into the mechanically competent core (Figure 1).

In continents, maximal T_e values in cratons are 110–150 km, while T_e distribution reveals complex "bimodal" behavior (clustering around 25–30 and 70 km), which does not allow for geometrical interpretation of T_e. This complexity is explained by *rheological stratification* of continental plates, specifically of their 35–70 km thick crust (see entry *Lithosphere, Continental*). Continental plates consist of several layers of contrasting strength (upper, lower, intermediate crust, mantle) that can be mechanically coupled or uncoupled. For plates younger than 200–400 Ma, T_e grows with age yet exhibiting large scatter of values between 0 and 70 km. Older plates (>500–700 Ma) are in stationary thermal regime; the age dependence of T_e is expectedly small, but T_e scatter is still strong revealing influence of factors other than thermal age. Using rock mechanics data (Figure 4), it can be shown that continental T_e should be strongly controlled by the crustal thickness and mechanical state of crust-mantle interface (Burov and Diament, 1995). When the lower crust is mechanically strong, crust and mantle are mechanically coupled yielding a single strong layer with high T_e:

$$T_e \approx h_1 + h_2 \ldots = \sum_n h_i \tag{9}$$

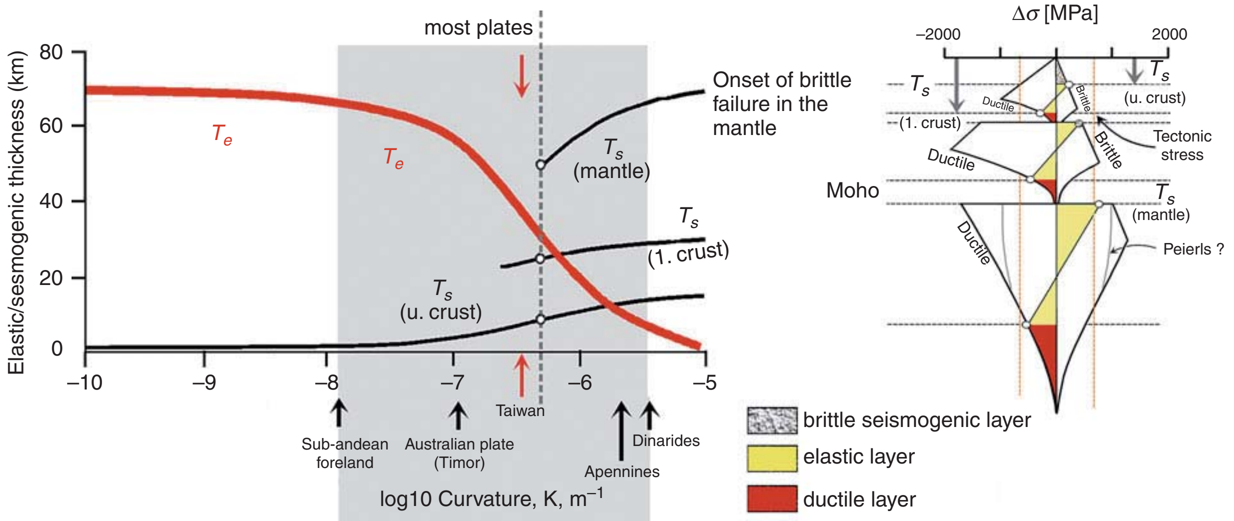
where h_i are thicknesses of crustal and mantle competent layers. Mechanical decoupling between crust and mantle leads to *structural weakening*, that is, dramatically smaller T_e (Burov and Diament, 1995):

$$T_e \approx \left(h_1^3 + h_2^3 \ldots\right)^{1/3} = \sqrt[3]{\sum_n h_i^3} \approx \max(h_i) < \sum_n h_i \tag{10}$$

"Decoupled" T_e roughly equals to the *mechanical thickness* of strongest layer in the lithosphere. In particular, $T_e \geq h_c$ (h_c is crustal thickness) identifies mantle as the strongest layer. Based on T_e data it has been shown that continental crust is systematically mechanically decoupled from mantle, except for cratons, and that mantle lithosphere is generally stronger or at least as strong as the crust (Burov and Diament, 1995). *Numerical thermo-mechanical models* have later confirmed these assertions (e.g., Burov, 2007, 2010).

Mechanical properties of the lithosphere and four modes of horizontal deformation

In addition to flexure under normal loads, four modes of lithosphere deformation and their combinations can develop under extensional or compressional conditions: (1) *pure shear* (extension: *McKenzie rifting model*; compression: plate *collision*), (2) *simple shear* (extension: *Wernicke rifting model*; compression: *subduction*), (3) *tensional/compressional instability* (extension: *boudinage*; compression: *folding-buckling*), and (4) *gravitationally instable pure shear* (slow *passive margins*, *Rayleigh-Taylor instabilities* due to thickening of a colder and thus denser lithosphere mantle overlying hotter and thus lighter asthenosphere). Ductile properties largely control these



Lithosphere, Mechanical Properties, Figure 4 Mechanical and rheological interpretation of continental T_e and *seismogenic layer thickness* T_s (*right*) computed for a representative YSE as function of local plate curvature K (*left*). After Burov (2010). Homogeneous horizontal stresses will shift T_e and T_s up or down in opposite directions, but will not change the character of their dependence on plate curvature.

deformation modes, including brittle fault spacing observed at surface (e.g., slow oceanic spreading centers vs. fast spreading centers). Brittle properties play an important role as well, specifically in localization of small and large-scale deformation, dyking, stabilization of cratons, transmission of tectonic stresses, and seismicity. Some of brittle mechanisms such as *low-angle faulting*, localization of large-scale *strike-slip* or *transform faults* and large-scale *dyke propagation* remain enigmatic.

Rheological stratification of continents is likely to influence the response of the lithosphere in all deformation modes. *Crust-mantle coupling* results in formation of narrow deep rifts while decoupling at different crustal levels may produce a variety of basins, from shallow large basins to narrow rifts. Under compression, *crust-mantle decoupling* plays a major role for continental subduction, UHP rock exhumation, slab detachment, formation of plateaus, or *lithosphere folding*. Lateral lower crustal flow is believed to be one of the key mechanisms affecting orogenic building and rift/passive margin formation or assuring feedback between surface and subsurface processes. For *mantle-continental lithosphere interactions*, such as *plume impingement*, the presence of ductile lower crust results in damping of the long-wavelength *dynamic topography* at surface, which is replaced by short-wavelength tectonic-scale deformation.

Uncertainties of data on the mechanical properties of the lithosphere

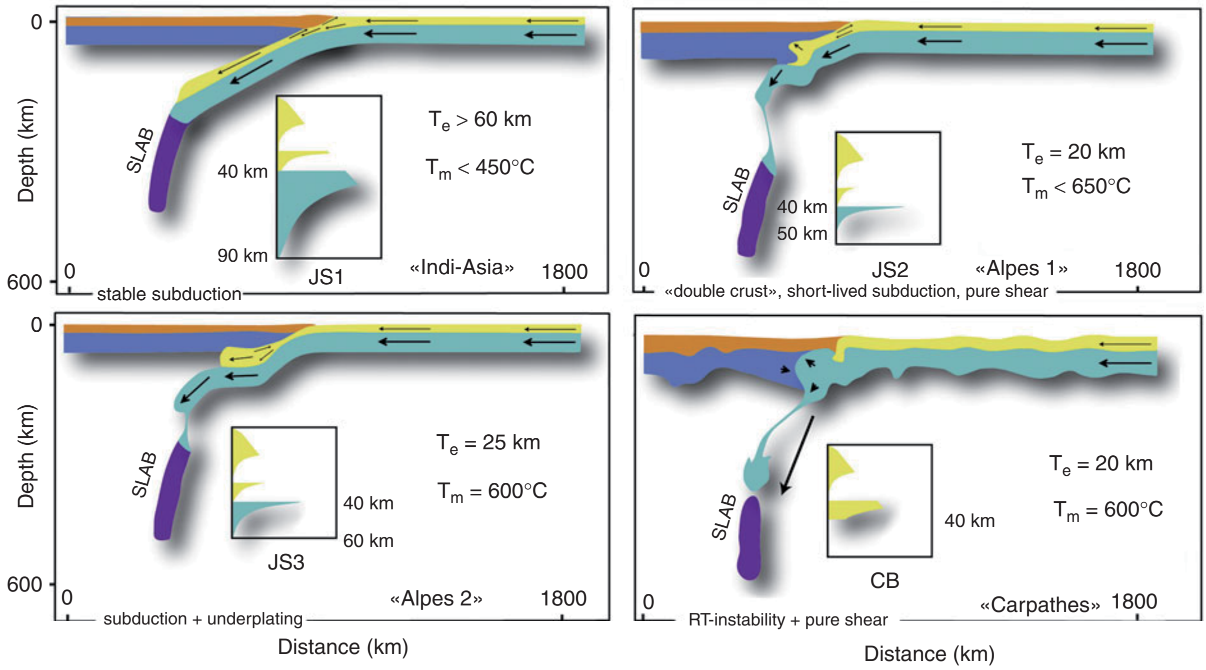
T_e studies play a key role in interpretation of *rock mechanics data*, allowing for parameterization of laboratory-based *flow laws* to geological time and spatial scales. Indeed, due to the large differences between time (10 orders) and spatial scales (6 orders) of the laboratory experiments and nature, *rock mechanics data* cannot be solely used to quantify the long-term *mechanical properties* of the lithosphere. Yet, since the *integrated strength B* of the lithosphere is directly related to T_e, one can use inelastic *flexural models* in combination with *rock mechanics data* (Equations 1–10) to constrain, parameterize, and validate rheology parameters for geological timescales. In particular, one can derive YSEs that are compatible with the host of different observations such as T_e, heat flow, seismic imagery, gravity, and so on. There is generally a good correlation between the observed T_e and T_e predicted from experimental YSE (Watts, 2001, Figure 4). In oceans, T_e values scale well with *rock mechanics data* and *thermal models of the lithosphere* (e.g., McNutt and Menard, 1982, Figure 3). In continents, T_e rheology parameterization is delicate but can be done if one respects the physics of underlying processes (Burov and Diament, 1995; Burov, 2010). Three types of *rheological envelopes* can be proposed for continental lithosphere: *Cratonic Jelly Sandwich*, *Jelly Sandwich*, *Crème brûlée* (with its variant hydrated *Banana Split* rheology) (Watts and Burov, 2003; Burov, 2010, Figure 2). The first two types evoke strong mantle lithosphere that can be either mechanically coupled with the crust (*Cratonic type*) or decoupled from it by weak lower or intermediate crustal layer (pure *Jelly Sandwich*). These first two rheologies dominate and correspond to 80–90% of all cases, as confirmed by matching T_e predicted from YSEs with the observed T_e and mechanical models. The *Crème brûlée* rheology infers weak lithosphere mantle, and applies to few cases of hot lithospheres (e.g., *Basins and Range Province*).

Mechanical properties of the lithosphere and styles of tectonic deformation

Important constraints on lithosphere rheology and properties come from physical considerations including *thermo-mechanical numerical models* of geodynamic processes. One can verify the validity of rheological assumptions by comparing predictions of thermo-mechanical models based on these assumptions with the observed tectonic evolution in various contexts such as collision and mountain belts (Figure 5), rifts and passive margins, and subduction zones and lithosphere folding. The analytical and *numerical thermo-mechanical models* of geological processes show that predicted evolution is highly sensitive to rheology choices (Figure 5). It has been demonstrated that in most cases mantle lithosphere has to be rheologically strong to insure stability of geological structures and the observed styles of deformation. In particular, stability of mountain ranges, continental *subduction* (e.g., Figure 5, Burov, 2010) and most observed *rifting modes* (Buck, 1991; Bassi, 1995) need initially strong mantle lithosphere with *mechanical thickness of the mantle part* more than 30–50 km. Collision models (Figure 5) quantify relationships between the collision styles and deep lithosphere structure and rheology. They show that tectonic evolution depends not only on the integrated strength (T_e) but as much on strength distribution with depth, that is, on which particular lithological layer (upper crust, lower crust, or mantle lithosphere) provides main contribution to T_e. Rift models show that *narrow rifts* are associated with initially strong mantle lithosphere, while metamorphic core complexes "need" warm lithospheres with negligible mantle strength. Wavelengths of *boudinage* and deca-kilometric upper crustal fault spacing observed in rift zones also appear to be a direct function of the ductile properties of the underlying lower crust. In compressional settings, observed *folding wavelengths* provide constraints on plate rheology since these wavelengths (50–1,000 km) are proportional to 5–10 times thickness of competent layers in the lithosphere and correlate with T_e values (e.g., Burov, 2007).

Links between different timescales: Burger's rheology model

The abundance of short-timescale observations such as *earthquake distributions* or *post-seismic relaxation* explains attempts to interpret these data in terms of long-term rheology. Yet, these attempts are typically based on



Lithosphere, Mechanical Properties, Figure 5 Numerical thermo-mechanical models of long-term geological processes (here, continental collision) allow for validation of inferred rheology profiles (*Jelly Sandwich*, JS1–JS3, and *Crème Brûlée*, CB) by testing the compatibility of model-predicted and observed tectonic evolution. Models JS2, JS3, and CB have similar T_e (20–25 km) but different mantle strength, leading to strong differences in resulting collision styles. The models show that strong mantle lithosphere is needed to drive continental subduction while strong crust cannot play same role (its low density prohibits subduction if there is no mantle drag). T_m – temperature at Moho. Colors: *orange*, *yellow* – crust; *blue*, *azure* – mantle; *purple* – oceanic slab. Modified from Burov (2010).

misinterpretation of Goetze's YSEs that are valid only for geodynamic strain rates. Earthquakes and relaxation occur at locally and temporarily high strain rates that are not representative of long-term rates. Hence, the fact that continental lithosphere below Moho depth is mainly aseismic cannot be interpreted as a sign of weak ductile behavior (Figure 4). It rather indicates that either mantle stress levels are insufficient to induce brittle sliding, or that seismogenic *Byerlee's law* does not operate at high pressure (>36–70 km depth), or that the frictional behavior is strain-rate dependent. At seismic strain rates (10^{16} times higher than geodynamic strain rates), the entire lithosphere acts as a brittle-elastic body and no ductile flow can occur. Strictly speaking, T_e and *maximal seismogenic depth* T_s should rather anti-correlate (Watts and Burov, 2003, Figure 4). However, T_s cannot be used to constrain T_e without knowing intraplate stress level. The fact that T_s distributions in oceans and continents are similar, while their rheological and mineralogical structures are different, adds to the argument that seismicity is primarily related to stress levels.

Post-seismic relaxation data provide controversial results yielding effective viscosities of deforming domains about 1–2 orders of magnitude smaller than *postglacial rebound data* that provide minimal viscosity (10^{19} to 5×10^{19} Pa s) of the Earth's weakest layer – asthenosphere. Since these estimates are based on inversion of *surface deformation*, they strongly depend on initial assumptions on lithosphere structure and properties, while it is impossible to determine which layer in the lithosphere relaxes post-seismic deformation. Nevertheless, whatever is this layer, the estimated viscosities appear too low for long-term properties. To explain this controversy, one can consider *Burger's model* of solids. According to this model, lithospheric behavior is described by two independent serially connected terms, one of which is *Kelvin solid* responsible for the primary creep (seismic, post-seismic) and the second one is *Maxwell solid* responsible for secondary long-term creep (geodynamic timescale). The viscosity of the first term is independent of that of the second term. In other words, the physics of deformation mechanisms activated at seismic-scale strain rates is different from that of the mechanisms acting at geodynamic strain rates.

Summary

Lithosphere has important long-term ductile, elastic, and brittle-plastic strength and is capable of maintaining (not relaxing) differential stresses at geological timescales. These properties can be accessed mainly from estimates of the *equivalent elastic thickness*, T_e, combined with *rock mechanics data*, validated by analytical and *numerical* *thermo-mechanical models* of geological and geodynamic processes. In oceans, T_e fits in the mechanical core of the lithosphere, but in continents, it generally does not represent any particular layer due to strong *rheological stratification*. The brittle-plastic properties of the lithosphere are governed by *Byerlee's law* to 30–40 km depth and likely by *Peierls* plasticity and/or *GBS* creep below. The ductile properties are dominated by *dislocation power-law creep*, which is strongly nonlinear and rock-type dependent. Highest *integrated strength* ($T_e \sim$ 100–150 km) is detected in cratons (*Cratonic Jelly Sandwich rheology*) where strong crust is mechanically coupled with strong mantle. In warmer continental lithospheres, mechanical decoupling between crust and mantle leads to *structural weakening* resulting in ~50% reduction of T_e (*Jelly Sandwich rheology*). In most cases, mantle lithosphere provides main contribution to the integrated plate strength. The cases where mantle is weaker than crust (*Crème-Brûlée rheology*) refer to hot lithospheres such as metamorphic core complexes. There is probably no exploitable link between short-term deformation such as earthquake data or post-seismic relaxation and long-term properties of the lithosphere.

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Cross-references

Earth's Structure, Global
Gravity Anomalies, Interpretation
Isostasy
Lithosphere, Continental
Lithosphere, Continental: Thermal Structure
Lithosphere, Oceanic: Thermal Structure
Seismicity, Intraplate

LITHOSPHERE, OCEANIC

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Definition and introduction

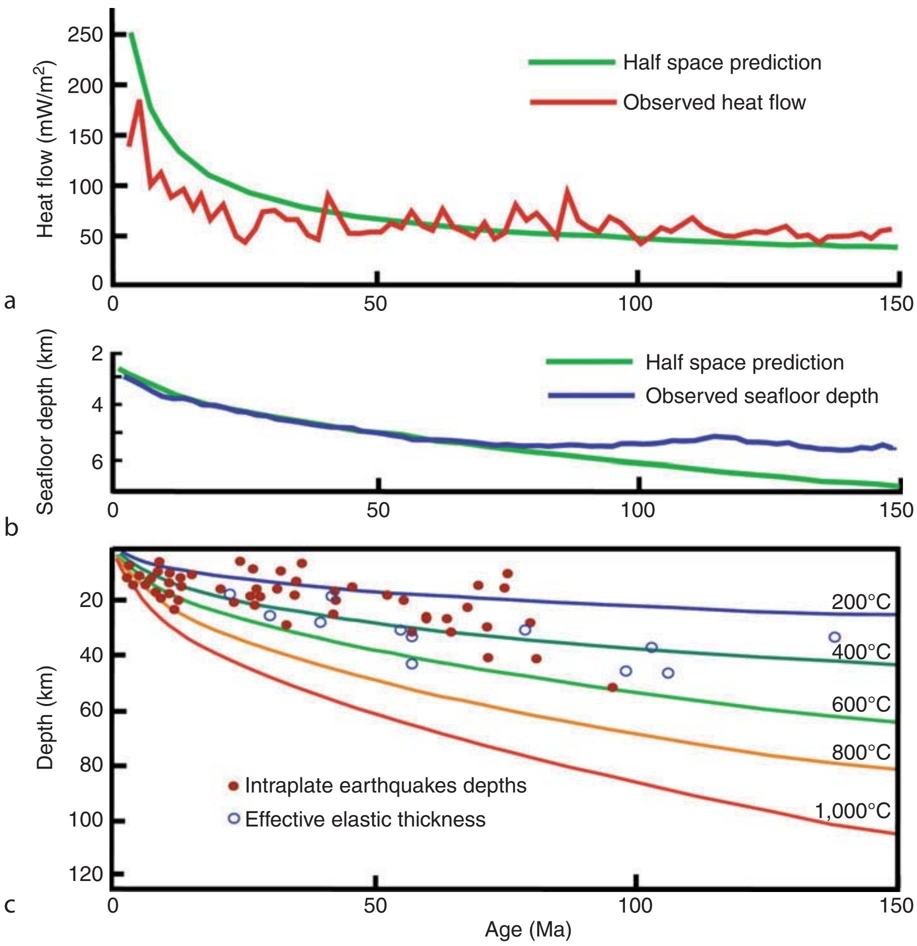
The term "lithosphere" is used or defined in a variety of formal and informal ways. All are fundamentally linked to a comparison of near-surface (lithosphere) mechanical and/or thermal properties with those properties deeper in the Earth (the asthenosphere). And all are linked to the concept of a thermal boundary layer between the bulk of the Earth and its atmosphere/ocean above. In this entry, we deal with the oceanic lithosphere, and we will define it as that part of the oceanic crust and the uppermost mantle that is usually created at mid-ocean spreading centers and that generally is consumed in subduction zones. For space purposes, we will specifically exclude oceanic sediments, island or continental arcs, and continental margins.

The ocean lithosphere is generally considered the outer carapace of the Earth that is cool, with an elastic rheology, and exhibiting brittle failure (e.g., failure by earthquake). This contrasts with the asthenosphere, which exhibits viscous rheology and higher temperatures. The concept of lithosphere was first developed in 1914 by Barrell (1914), who realized that the outer part of the Earth must be able to support tectonic loads over geological timescales. Since that time, workers additionally have defined the lithosphere in terms of plates that move coherently in plate tectonics. Alternatively, the oceanic lithosphere may be defined as the relatively high velocity or low attenuation "seismic lid," or as that part of the earth that can sustain earthquakes. These various definitions are not necessarily mutually exclusive, but it is important to consider how the term lithosphere is used (Figure 1).

The uppermost mantle lithosphere

The mantle component of the lithosphere is composed of peridotite, and the thermomechanical transition from lithosphere to asthenosphere is defined on the basis of the data used to detect that transition. Many have used the seismic low-velocity zone often inferred for the upper mantle as the essential (e.g., Forsyth, 1977; Nishimura and Forsyth, 1989). Typically exploiting surface wave observations, these studies have been very useful in defining variations in lithospheric thickness.

Another approach to estimating lithospheric thickness is derived from observations of the response of the lithosphere to tectonic loading (e.g., Calmant et al., 1990). Specifically, the addition of a large seamount or island should cause the lithosphere (including the seafloor) to deflect downward as the load is accommodated by the asthenosphere. The amount of deflection or flexure, and the regional distribution of the deflection, can be considered a function of the thickness of the strong lithospheric plate. The topography of the seafloor around a seamount can be used to calculate this "effective thickness." A compilation of thicknesses is shown in Figure 1c, and reveals the thickness of the lithosphere coincides with isotherms between 400°C and 600°C in the mantle. Similar values for lithospheric thickness are obtained if one uses the depths of intraplate earthquakes, which presumably are indicators of brittle failure confined to the lithosphere (Wiens and Stein, 1983) see (Figure 1c).



Lithosphere, Oceanic, Figure 1 Different observables are used to constrain models for the oceanic lithosphere, primarily its mantle component. (**a**) Observed heat flow data (in *red*) averaged in 2-million-year bins (Stein and Stein, 1992) compared to predictions for a conductively cooled half space (in *green*). Heat flow is too low and scattered for ages less about 40 Ma, indicating nonconductive (hydrothermal circulation) cooling must be taking place. (**b**) Observed seafloor depth (in *blue*) compared to predictions for a conductively cooled half space (in *green*). Depths are too shallow beyond 80 Ma, indicating temperatures have stabilized and heat is being supplied from below. (**c**) Plot of earthquakes depths and effective elastic thickness on top of expected isotherms for a half-space model.