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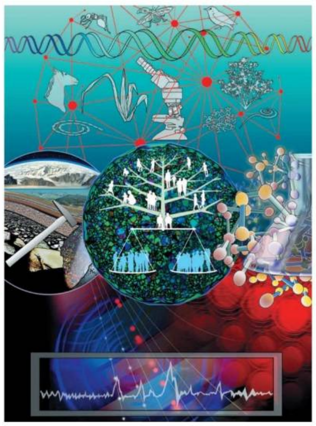

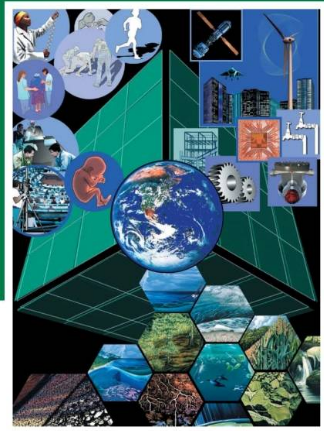

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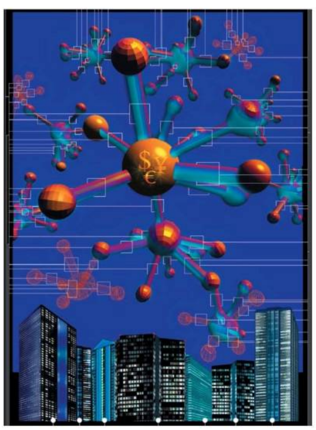

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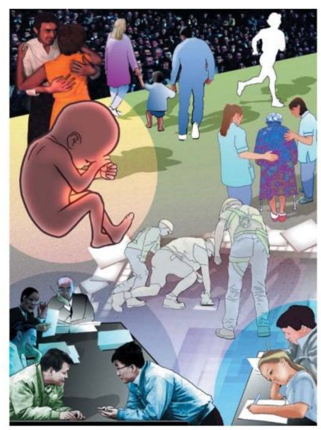

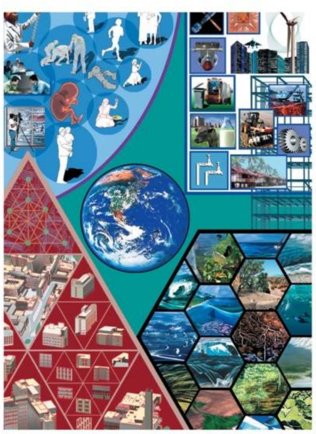

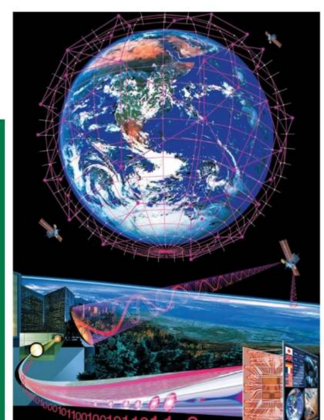

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EARTH AND ATMOSPHERIC SCIENCES

GEOLOGY

Edited by: Kurt Stüwe *Institut für Geologie und Paläontologie, Universitaet Graz,Austria*

Bernhard Grasemann *Department of Geological Sciences, University of Vienna,Austria*

Benedetto De Vivo *Dipartimento di Geofisica e Vulcanologia, Universita di Napoli "Federico II",Italy*

	Universität Potsdam,Germany
	George E. Hilley *Institut für Geowissenschaften, Universität Potsdam,Germany*
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TECTONICS AND GEODYNAMICS

Grasemann, Bernhard
University of Vienna, Vienna, Austria

Stüwe, Kurt
Karl Franzens University of Graz, Graz, Austria

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Contents

Summary

The science of geodynamics and tectonics includes the description and interpretation of a large variety of geological processes that operate in the earth. However, we generally think of geodynamic processes of those that act on the scale of the whole lithosphere. This contribution summarizes some fundamental concepts of geodynamics and tectonics, and serves as an introduction to articles on specific subjects within the topic. Among the methods of describing geodynamic processes we discern between energetic, kinematic and dynamic descriptions. Energetic descriptions are those that consider the production, distribution and redistribution of thermal energy in the lithosphere. Kinematic descriptions are those used to describe the movement and geometry of rocks units and surfaces. Dynamic descriptions are those that use force balance considerations to describe orogenic processes. The application of these methods to the description of geodynamic processes is illustrated simple model concepts of the rheology of the lithosphere and the principle of isostasy as examples. The contribution if rounded off with an outlook on the future of geodynamics.

1. Introduction

Geodynamics is the science describing the dynamic processes that govern the large scale structure of earth. Geodynamic processes have operated throughout the billions of years of the earth's history to create, destroy and recreate continents and oceans, geological provinces and terranes, mountain chains and basins, and all the mineral and hydrocarbon deposits so essential to our society. Thus, "geodynamic processes" are understood to include a large variety of processes and earth scientists use the term quite loosely. Nevertheless, when discussing "geodynamics and tectonics" we generally think of processes that act on the scale of lithospheric plates or plate boundaries, rather than on the scale of a single outcrop. In the past decade the term "geodynamics" has often been used as a fashionable synonym to "tectonics", which traditionally has only been understood to be the science of the kinematics of rocks on a large scale, e.g., in the context of terms like "thrust tectonics", "extensional tectonics", "subduction tectonics" and so on. However, geodynamics also includes the conceptual description of physical processes governing tectonics, the combination of thermal and mechanical descriptions for integrated physical interpretation of the earth and more. In general, it may be said that geodynamic processes are described using energetic, kinematic and dynamic descriptions. While these three methods of description cannot be separated strictly, they each use some characteristic variables: Energetic descriptions are involved with the distribution of thermal energy using variables like heat or temperature. Kinematic descriptions are those using parameters like velocity and strain. Dynamic descriptions are those using variables like stress and force. Each of these three descriptions is discussed in this contribution with some applications to some geodynamic processes on the lithospheric scale. We therefore begin with a brief summary of the concept of plate tectonics.

2. The Theory of Plate Tectonics

2.1. The Geodynamic Concept of Plate Tectonics

The earth's outermost rigid layer is called lithosphere and it consists of a rigid upper mantle and a crust. The crust may be either of oceanic or crustal type. The theory of plate tectonics states that this lithosphere is broken into 7 major plates (African, North American, South American, Eurasian, Australian, Antarctic, and Pacific plates) and several minor plates (e.g., Arabian, Nazca, Cocos, Juan-de-Fuca and Philippines plates) sliding over the plastic asthenosphere, which is the upper layer of the mantle (Figure 1). The lithospheric plates are all moving in different directions and at different velocities between 1 to more than 10 cm a^{-1} relative to each other. The place where the two plates meet is called a plate boundary. Where they interact, along these margins, important geological processes take place, such as the formation of mountain belts, earthquakes, and volcanoes. These plate boundaries have different names depending on how the two plates are moving in relationship to each other: (i) convergent, (ii) divergent and (iii) transform boundaries. Before discussing these plate boundaries in some more detail, we also note that some important mountain belts, for example the Tien Shan, have also formed inside continental plates, and that other important geological processes like Hot Spot volcanism also occur inside oceanic plates.

Figure 1. The intraplate stress field of the world superimposed on a rough plate tectonic subdivision of earth. Different symbols indicate different methods of stress determination including earthquake focal mechanisms, borehole breakouts and geological indicators. Different shadings indicate different deformation regimes: darkest are thrust faults, medium grey are normal faults and light shading are strike slip faults. This map was created using the CASMO facility on the world stress map project home page. The map is also modified after Stüwe (2002).

2.1.1. Convergent Boundaries:

If the size of the earth has not changed significantly during the last 200 Ma (which is about the age of the oldest oceanic lithosphere) this implies that the lithosphere must be destroyed at about the same rate as it is being created. This recycling of lithospheric material takes place along convergent boundaries or collision zones (destructive plate boundary). Collision can occur:

between an oceanic and a continental plate
between two oceanic plates, or:
between two continental plates.

Oceanic-continental convergence is associated with up to 10 km deep trenches, where the oceanic lithosphere is subducted beneath the continental plate (e.g., the Nazca Plate beneath the South American Plate). Subduction is frequently associated with eruptive volcanic activities where the magma is either generated by the partial melting of the subducted oceanic slab, or the overlying continental lithosphere, or both.

When two oceanic plates converge, one is subducted beneath the other, and similar to oceanic-continental subduction a deep trench is formed (e.g., Mariana Trench). Subduction processes in oceanic-oceanic plate convergence also result in the formation of volcanoes, which are called, because of their parallel arrangement to the generally curved trenches, island arcs.

Continental-continental convergence is the result of the collision of two continents forming a collision orogen (e.g., the Himalayas after collision of India with Asia some 55 Ma ago). Generally, the continental collision followed an oceanic-continental subduction zone. Such collisional orogens are characterized by fold-and-thrust belts, regional metamorphism, igneous activities, exhumation of high-grade and ultra-high-pressure rocks and surface uplift and erosion of huge rock masses.

2.1.2. Divergent Boundaries

Divergent plate boundaries occur along spreading centercenters where plates are moving apart and new lithosphere is created by magma pushing up from the mantle (constructive plate boundary). One of the best-known examples is the Mid-Atlantic Ridge, which is one segment of the global mid-ocean ridge system that encircles the earth and has a total length of some 60000 km. The rate of spreading along the Mid-Atlantic Ridge averages about 2.5 cm a^{-1}. This seafloor spreading over the past 200 Ma has caused the Atlantic Ocean to grow from narrow rift between the continents of Europe, Africa, and the Americas into the vast ocean that exists today. However, divergent plate boundaries like the Mid-Atlantic Ridge never start out as boundaries between two oceanic plates. They always commence as rifts in side the continents. The causes for such rifting remain unclear, but it is fiercely debated if they initiate in response to hot spots underneath the continents or in response to gravitational potential energy differences. A new spreading center probably forming the earth's next major ocean may be developing under Africa along the East African Rift Zone. A further stage of divergent plate boundary development may be observed in the Red Sea where the first oceanic lithosphere begins to form.

2.1.3. Transform Boundaries

The zone between two plates sliding horizontally past one another is called a transform-fault boundary, or simply a transform boundary. Transform faults connect either two spreading centercenters (divergent plate boundaries) or, less commonly, trenches (convergent plate boundaries). Therefore most transform faults are found on the ocean floor where they produce the offset appearance of mid ocean ridges. However, transform boundaries may also occur on land (e.g., San Andreas fault zone in California connecting two divergent boundaries).

It is important to note that not all plate boundaries on earth are as simple as the main types discussed above, especially if plate-movement deformation occurred over a long time span with changing kinematics and extends over a broad belt (i.e., plate-boundary zone). Such plate-boundary zones frequently involve at least two large plates and one or more micro-plates resulting in complicated geological structures and earthquake patterns (e.g., Mediterranean-Alpine region between the Eurasian and African Plates).

2.2. Global Seismicity in the Concept of Plate Tectonics

The theory of plate tectonics furthermore explains four types of seismic zones:

Seismic zones along rift systems
Seismic zones along transform boundaries
Seismic zones related to subduction zones
Seismic zones associated with collision tectonics

The first type of seismic zone follows the mid ocean rift system and is associated with the volcanic activity along the axis of the ridges (e.g., Island). The seismic activity is generally low, and it occurs at very shallow depths because the lithosphere is very thin and weak at these divergent boundaries.

The second type of seismic zone is associated with transform plate boundaries or large strike slip faults causing friction between neighboring plates (e.g., North Anatolian Fault or San Andreas Fault). Earthquakes are shallow-focus events generally without any volcanic activity. As large strike slip faults are always bend and segmented in several sections activity does not always occur along the entire length of the fault during any one earthquake.

The third type of earthquake is related to subduction zones along convergent plate margins. One plate is thrust or subducted under the other plate so that a deep ocean trench is produced (e.g., Peru - Chile trench, where the Pacific plate is being subducted under the South American plate). Earthquakes associated with subduction zones can be shallow, intermediate, or deep, according to its location on the down going lithospheric slab (Wadati-Benioff zone).

The fourth type of seismic zone is associated with collision tectonics where shallow earthquakes are associated with intense shortening and formation of high mountain ranges (e.g., broad swath of seismicity from Burma to the Mediterranean, crossing the Himalayas, Iran, Turkey, to Gilbraltar).

Numerous processes, such as for example, climate change, earthquakes, volcanism or erosion are of major concern for human life. The motion at plate tectonic boundaries is only in the order of a few mm – cm a^{-1} and can therefore only be observed by means of high-resolution geodetic methods over a time span of years (e.g., Global Positioning System). However, within seconds an earthquake or volcanic eruption can unleash bursts of energy far more powerful than anything we can generate. Besides all these hazards, people also benefit from the forces and consequences caused by plate tectonics (e.g., geothermal energy, generation of natural resources).

Due to the importance of plate tectonic processes for our life, separate articles of this encyclopedia are dedicated to the plate tectonic processes: *Plate Tectonics of Continents and Oceans* by Martin Meschede (Ernst-Moritz-Arndt-Universität Greifswald, Germany), which covers many aspects of the earth's composition and age, plates, plate boundaries, triple junctions and hot spots;*Neotectonics* by Manfred Strecker and co-workers (Potsdam University, Germany) which emphasizes all features of active tectonic processes including geomorphological, paleoseismological, geodetical and geophysical methods. A special emphasis is given on neotectonic movements and climate patterns.

3. The Description of Geodynamic Processes

The following sections discuss some fundamental concepts in geodynamics and serve as a basis for other articles covering important subjects related to the dynamics of the earth. Excellent, detailed mathematical and physical introductions to this topic are given in the Bibliography. *Geodynamics of compressional orogens* by Jean Braun (Australian National University, Australia) shows an application of the fundamentals of geodynamics by demonstrating the use of complex numerical methods in order to investigate the geodynamic evolutions and processes of past and present orogens.

3.1. Energetic Descriptions of Geodynamic Processes

An enormous number of rock properties that may be observed in the field are a strong function of temperature. This includes, for example, the type of metamorphic paragenesis that a given rock contains or the mechanism by which it deforms. Also, thermal energy is ultimately the driving force of all plate tectonic processes and energetic descriptions of geodynamic processes are therefore amongst the most fundamental and important methods of description. Finally, the deformation of rocks may also be viewed as an energetic process, as the work done on a rock by deforming it is stored as energy. The thermal energy content of a rock in Joule per cubic meter is readily converted into temperature by dividing it by its density (in kg m^{-3}) and heat capacity (in J kg^{-1} K^{-1}). Thermal energy is produced and distributed in the lithosphere by three processes namely conduction, convection and production.

3.1.1. Conductive Heat Transfer

Heat conduction is a diffusive process, where molecules transmit their kinetic energy to other molecules by colliding with them. The conduction of thermal energy is described by the famous diffusion equation stating that the rate of temperature change of a given point (rock) is proportional to the spatial curvature of the temperature profile around this point. The proportionality constant is called the thermal diffusivity and is given by the ratio of thermal conductivity, density and heat capacity. Heat conduction is one of the most important processes governing the thermal structure of the lithosphere (for example by conductive heat loss to the surface, by regional metamorphism etc.) and an enormous number of studies have employed numerical and analytical solutions of the diffusion equation to describe such processes. Probably the most successful conductive model of all time is known as the *age-depth-heat flow relationship of the oceanic lithosphere*. This model describes the thermal structure of the oceanic lithosphere with an extremely simplistic diffusion model which has successfully described a number of geological observations that may independently be checked, namely the water depth of the oceans, the age of the oceanic lithosphere and the surface heat flow in the oceanic lithosphere, all as a function of distance form the mid oceanic ridges.

While thermal modeling has solved an uncountable number of geological problems, many heat conduction problems may simply be solved by considering a simple proportionality between the length scale of a heat conduction process L, the time scale of the conductive process τ. This relationship may be expressed as:

$\tau \sim L^2/\kappa$ (1)

where κ is the thermal diffusivity which is roughly $\kappa=10^{-6}$ m^2s^{-1} in rocks and varies by rarely more than a factor of 2 around this value. Eq. (1) can readily be used to estimate characteristic time and length scales of heat conduction processes in the lithosphere. For example, the duration of a regional metamorphic event affecting a crust of some 30 km thickness is readily estimated to be some 30 Ma or, more conservatively, "some tens of millions of years". Correspondingly, the duration to thermally equilibrate a soft-boiled egg with a length scale of several centimeters is about 10 minutes. As such, equation (1) can be used to estimate the duration of cooling of magmatic intrusions, the size of a metamorphic terrain for which the duration of metamorphism is known and countless other geodynamic processes.

3.1.2. Convective Heat Transfer

Convection is the process of redistributing thermal energy by mass transfer. Convection is best known to us from the convection of the upper mantle as the driving force of hot spots, and ultimately plate tectonics. Convective processes are generally thought of to be circulating, while the term *advection* is generally used when the material transport is in one direction only. Intrusion of magmatic plutons from the base of the crust into higher crustal levels is a typical example of advective heat transport. Advection of heat may occur by three different media:

- advection of heat by fluids
- advection of heat by magma (for example intrusion or volcanism)
- advection of heat by the movement of rock (for example erosion or deformation)

The importance of heat *advection by metamorphic fluids* is now known to be of negligible influence to the thermal energy budget of metamorphic terrains as a whole. However, the other two transport mechanisms are extremely significant.

Advection of heat by magma is responsible for all kinds of contact metamorphism, including low-P high -T metamorphic terrains, which are characterized by peak metamorphic PT ratios which cannot be interpreted otherwise. The advective redistribution of heat from the lower crust to the middle crust is also extremely significant for the mechanical and mechanical evolution of the lithosphere as it can substantially decrease the mean rheological strength of the crust.

Advection of heat by solid state motion, for example by deformation, is responsible for inverted metamorphic gradients when warm nappes are thrust over colder rock units. Erosion may also act as an advection agent as it causes the crust to exhume. Recent research has shown that rapid erosion may cause advective heating and therefore weakening of the crust that is significant enough so that deformation events are triggered.

Figure 2. Modeled exhumation (temperature-depth) path of rocks exhuming from

10-100 km depth at (a) 1 mm/a and (b) 5 mm/a. Note that higher exhumation rates favour isothermal decompression of rocks (modified after Grasemann *et al.*, 1998).

A useful relationship to estimate the importance of advection processes is the thermal Peclet number *Pe* defined as:

$$Pe = uL/\kappa \qquad (2)$$

where u is the advection rate in meters per second, L is the length scale of the advection process and κ is the thermal diffusivity as in equation (1). Thus, Pe is dimensionless. Thermal Peclet numbers much smaller than 1 indicate that advection is insignificant and is outweighed by diffusive dissipation. If Pe is much larger than 1 advection is more important than diffusion and if it is around 1, both processes need considering. For example, if the crust with a length scale of 30 km erodes at the surface with 1 kilometer per million years, then Pe is of the order of 1 indicating the both diffusive and advective processes are relevant during regional metamorphism of orogenic belts. We see evidence for this in the perfectly smooth curves of modeled PT paths from regional metamorphic terrains indicating how conductive heating and advection by erosion interplay (Figure 2).

3.1.3. Heat Production

Thermal energy may not only be redistributed in the lithosphere, it may also be newly produced or consumed during geodynamic processes. Three thermal energy sources may be discerned:

- radioactive heat production
- chemical heat production
- mechanical heat production.

Radioactive heat production is an important contributor to the thermal energy budget of metamorphic terrains and the earth as a whole. It is partially because of the unknown radioactivity that Lord Kelvin underestimated the age of earth based on surface heat flow measurements by 2 orders of magnitude (interestingly, he also underestimated the age because he neglected the advection of heat to the base of the lithosphere in the mantle, although the knowledge of convection in the upper mantle was well established at his time). Radioactive heat production in the crust is of the order of some Milli-Watts per cubic meter, which - in the absence of heat conduction - would translate to some tens of degrees heating per million years. Radioactive heat production in the mantle is negligible.

Chemical heat production is threefold:

- Chemical heat production due to phase changes
- Chemical heat production due to dehydration reaction
- Chemical heat production due to solid - solid reaction.

Chemical heat production due to all solid state reactions during metamorphism is of negligible influence to the crustal heat budget, and may only play a role to very local thermal buffering processes on the grain scale. However, chemical heat production due to phase changes is extremely significant. While the heat changes during condensation/evaporation reactions are to us an everyday example when we step out of the shower, their geodynamic relevance is restricted to hydrothermal settings. However, solid-liquid phase transformations are familiar to us from all partially melted metamorphic terrains and the heat changes associated with this process are extremely relevant to the thermal evolution of such terrains. The heat associated with crystallization is of the order of 300 kJ per kilogram of melted rock.

Mechanical heat production, also known as frictional heating or viscous dissipation, is the largest unknown of all heat sources in the lithosphere. We know of the geological existence from pseudotachylites and other local observations, but its relevance to the heat budget of regional metamorphic terrains still remains one of the most debated topics in the earth sciences.

As modeling of heat transfer processes is closely related to geochronological methods, *Geochronology and Isotope Geology* by Igor Villa (Bern University, Switzerland) highlights the fast evolving geochronological methods, their perspectives and limitations.

3.2. Kinematic Description of Geodynamic Processes

Many of our geological observations in the field are spatial observations, i.e., the geometry and location and position of a given rock or rock unit. From such observations we infer the rock kinematics, for example, the shear sense on a shear zone. Entire sub disciplines of geology are concerned with spatial and kinematic descriptions, for example balancing cross sections, large parts of geomorphology and structural geology or paleogeography. Deformation and displacement are described using the deformation tensor, which is briefly discussed in Section 3.3. However, the most crucial point of any kinematic description is the correct reference frame.

3.2.1. Vertical Reference Frames

The confusion about vertical motions in the lithosphere is a good example for the need of precision with reference frames. A classical example concerns the evidence of raised beaches relative to sea level. Is this evidence for a raising continent or a decreasing sea level? In order to solve this problem it is necessary to define some reference frame. An absolute framework is given by the geoid. The geoid is the surface of constant gravitational potential energy and geoid anomalies are the distance of this surface from a reference ellipsoid that is used for cartography (Figure 3). Sea level changes that occur relative to the geoid are referred to as *eustatic* sea level changes and may be interpreted in terms of changes of the water volume in the oceans.

Figure 3. Geoid map of the world. Modified after Stüwe (2002).

Another important example for the need of care with reference frames is the different use of reference frames between metamorphic petrologists and geomorphologist. Metamorphic petrologist use the surface as the vertical zero and measure distances in the crust positively *downwards*, e.g., a rock is referred to as being 15 km in the crust, meaning ”from the surface”. In contrast, geomorphologists use sea level or the geoid as a vertical zero and measure distances positively *upwards*, e.g., a mountain is referred to as being 5 km high. Thus, studies that attempt to couple tectonic or metamorphic information on the depth of rocks in the crust with topographic information on the elevation of the surface see themselves faced with the need to couple different reference frames. For this, it is useful to discern between Lagrangian and Eulerian reference frames.

3.2.2. Lagrangian and Eulerian Reference Frames

Lagrangian reference frames are reference frames that are attached to a moving or deforming material. Thus, Lagrangian reference frames are also referred to as ”material frame” or ”material description”. For example, if we consider the metamorphic evolution of a rock, it is useful to fix the coordinate system to the rocks and observe the changes of pressure and temperature of a constant coordinate that moves with the particles. In contrast, a Eulerian description, (also referred to as a ”spatial description”) is fixed to an external reference frame. Within a Eulerian description, the evolution of rocks would track through the grid. Both reference frames have their advantages and disadvantages. The principle advantage of Eulerian descriptions is that a numerical grid has to be set up once, but can remained unchanged through the modeling process. The principle advantage of Lagrangian grids is that particle points remain attached to the same nodes and it is therefore straightforward to observe the evolution of single rocks.

3.3. Dynamic description of geodynamic processes

Plate tectonics is a consequence of forces acting on the lithosphere. These forces ultimately arise from the thermal convection in the mantle, but more directly they are largely the consequence of potential energy differences in the lithosphere itself. The principles of the stress and deformation tensors are fundamental to the understanding of dynamic and kinematic descriptions and we therefore here provide a brief overview.

3.3.1. Stress and Deformation

Stress is conventionally defined as a force acting on some area, but this physical quantity is more accurately termed "traction." A traction is a force per unit area acting on a specified surface. This traction can be resolved into its two vector components: (i) The first component act normal (perpendicular) to the surface (normal traction). (ii) The second component acts parallel to the surface (shear traction). "Stress" strictly speaking, refers to the whole collection of tractions acting on each and every plane of every conceivable orientation passing through a discrete point in a body at a given instant of time. This state of stress is represented by an ellipsoid named the stress ellipsoid, which can be mathematically described by the stress tensor of the second rank. This tensor can be defined by three pairs of normal and shear stresses acting on three arbitrary oriented orthogonal planes. A large variety of geological problems may be sufficiently described with the eigenvalues of the stress tensor. These eigenvalues are termed the principle stresses. Three principal stress components $_1$, $_2$, and $_3$ act on three orthogonal planes, which are eigenvalues of $_{ij}$. The principal stresses have by definition no shear components and correspond to the major axes of the stress ellipsoid. These directions can be easily derived using the graphical tool of the Mohr circle.

Rotation of a symmetric second-rank tensor about one of its principal axes has a geometrical analog (Figure 4a), called the Mohr Circle named after its discoverer Otto Christian Mohr (German civil engineer, 1882). In Mohr space, the diagonal components of the tensor are plotted along the horizontal axis (i.e., normal stresses); non-diagonal components plot along the vertical axis (i.e., shear stresses). The Mohr Circle is defined by its center and radius. The center, which always lies on the horizontal axis, is simply the sum average of the orthogonal normal stresses. The radius of the circle is defined by the maximum shear stress. Any point on the circle defines the normal and shear stresses on a plane and from the geometry of the circle we can determine the angle the principal axis makes with respect to the normal to the plane.

Figure 4 (a) Mohr Circle construction of a symmetric tensor *S*.
(b) Deformation tensor *D* relating two opposite corners of a unit square with the same corners of its deformed equivalent

Stresses cause solids to deform. At a scale at which deformation can be approximated to be continuous and homogeneous the deformation can be characterized an ellipsoid called the *finite strain ellipsoid* representing a deformed unit sphere. Mathematically this deformation can be described by the deformation tensor ***D***. In two-dimensions ***D*** relates two opposite corners of a unit square with the same corners of its deformed equivalent (Figure 4b). Contrary to the stress tensor, where only six independent components are necessary to define the state of stress at a point, ***D*** needs nine components representing axial length and orientations of the three principal axes of the finite strain ellipsoid and the rotation in space made by these axes. The exact shape of any ellipsoid is fully characterized by only two ratios between the longest and intermediate axis (l_1/l_2) and between the intermediate and shortest axis (l_2/l_3). In a graph where the first ratio is plotted against the other, one can create a space where every ellipsoid can be plotted. This space is called the Flinn diagram. Flat (pancake-shaped) ellipsoids plot along the horizontal axis and long (cigar-shaped) ellipsoids plot near the vertical axis where strain is said to be constrictional. The diagonal ($l_1/l_2 = l_2/l_3$) defines the domain of plane strain deformation where the axis l_2 axis of finite strain field ellipsoid keeps constant during deformation (i.e., l_2 is a direction on non-deformation).

General deformation has two end-members referred to as simple and pure shear. Simple shear is analogous to the process that occurs when a deck of cards is sheared to the right or left with each successively intervening card sliding over its lower neighbor. A cube subjected to a simple shear event is converted into a parallelogram resulting in a rotation of the finite strain axes. The sides of the parallelogram will progressively lengthen as deformation proceeds, but the top and bottom surfaces neither stretch nor shorten. Instead they maintain their original length, which is the length of the edge of the original cube. In contrast to simple shear, pure shear transforms a square into a rectangle by homogeneous flattening. One side of the square is shortened in one direction and elongated in the other direction. With pure shear the sides of the square remain parallel and perpendicular after the deformation event.

A more elaborate approach, including some aspects of continuum mechanics, and also perspectives of analytical, physical and numerical modeling of deformation processes using different (and sometimes very complex) rheologies is given in*Structural Geology* by Neil Mancktelow (ETH Zürich, Switzerland).

3.3.2. Deformation Mechanisms

While it is evident that there is no deformation without stress, their actual relationship is not trivial to define on a physical basis. In other words, stating that stress and deformation in rocks are related is quite a different matter from physically determining the extent of their actual relationship(s). In materials science and geology, the term rheology is used to describe the ability of materials under stress to deform or to flow. Plate tectonics, as manifest by mountain building, earthquakes and volcanoes occur by plastic or brittle deformation of the rocks and minerals that comprise the oceanic and continental lithosphere. Temperature, pressure and rate of deformation to a large extent define the nature of deformation for most minerals and rocks in the interior of the earth. However, the chemical environment (notably water and oxygen fugacity, activity of silica) may also have a significant influence. Therefore, insight into the mechanisms by which rocks and minerals deform under conditions that approximate those in the earth is critical to our understanding of the processes that shape our planet. There are a wide variety of deformation mechanisms that are of particular interest to structural geology. However, when dealing with geodynamic on the lithospheric scale, it is usually sufficient to summarize deformation mechanisms under three different simple mechanisms:

Brittle deformation
Elastic deformation
Viscous deformation

Brittle deformation is strictly not a deformation mechanism, as it does not place stress and strain in a relationship to each other. Rather, it is a law expressing a stress state, namely that at which fracture occurs. This stress state is usually described using the concept of Coulomb's friction law (named after the French physicist Charles-Augustin de Coulomb, 1736-1806) for pre-existing fault surfaces. This law predicts a linear increase of rock strength with normal stresses acting on the rock. In the crust, Byerlee's friction law (empirically derived from experimental determination of "maximum shear stress" on a wide range of rock types) is the quantitative relationship used to express this law. The deformation law used to describe the deformation of brittle materials is usually plastic deformation which states that a certain threshold stress is required for deformation to commence but is thereafter independent of strain or strain rate.

The strength increase with depth differs between extensional, strike-slip, and compressional situations, due to the different orientation of fractures and normal stresses on them. Apparently, the frictional strength is independent of mineralogy, strain rate, and temperature. However, each of these parameters may have some effect. The presence of fluid pressure will tend to decrease the strength of the fault.

Elastic deformation law states that stress and strain are proportional to each other. In the most simple case, this relationship is linear and is then referred to as Hook's law. The proportionality constant between stress and strain is termed Young's module. Elastic deformation can be observed on the outcrop scale through the distribution of fracture, but it is more apparently manifest on the plate scale, for example in flexural foreland basins or subduction zones. The accumulation of elastic strains can also be measured with modern GPS methods and has become an important tool in the prediction of earthquakes.

Viscous deformation dominates the lower crust and most of the mantle part of the lithosphere and is the most widely used deformation mechanism to describe plate tectonic deformation at the largest scale. Viscous deformation is also loosely referred to as "ductile" or "plastic" deformation, but these terms describe different processes: Ductile deformation is a more general term summarizing all non-brittle deformation processes without defining a specific constitutive relationship. Plastic deformation on the other hand is a well-defined description of a ductile deformation mechanism, where strain rate is independent of stress. Thus, the term plastic deformation should only be used to describe exactly that.

Typical plastically deforming materials are sand and all other fractured materials. Thus, plastic rheologies are used by engineering geologists. Viscous deformation is described in its most simple case by a direct linear proportionality between stress and strain rate. The proportionality constant is called viscosity. Such a linear relationship is called Newtonian flow. In geology it is complicated by two important facts: (i) Viscosity is a strong exponential function of temperature (this function is called the Arrhenius relationship named after the Swedish physicist Svante August Arrhenius, 1859-1927). (ii) In rocks, the proportionality between stress and strain rate is typically not linear, but governed by a stress exponent stating that stress raised to some power is proportional to strain rate. The stress exponent of rocks is typically between 3 and 5, so that the application of a doubled stress results in an 8 to 32 fold increase in strain rate.

4. Important Concepts in Geodynamics and Tectonics

4.1. Rheology of the Lithosphere

On the scale of entire lithospheric plates, the rheology of the lithosphere has simply been assumed to be viscous. In order to avoid the depth dependent description of rheology, the assumptions of plain strain or thin viscous sheet deformation have been made for plan view descriptions. However, a range of observations show that the rheology must change with depth. For example, we observe that rocks from shallow crustal levels have deformed in a brittle fashion, while rocks from lower crustal levels deform in a ductile manner. Also, the distribution of seismicity on the globe shows that oceanic lithosphere is stronger than continental lithosphere, although it is substantially thinner. In order to explain such observations, there is the need for more refined models for the rheology of the lithosphere. Such a model is readily developed using some intuitive relationships on the principle constitutive relationships that govern the lithospheric rheology. The model discussed on the following is based on considerations of Brace and Goetze around 1970 and has received the name ”Brace-Goetze lithosphere”.

According to Byerlee's law, the brittle strength of rocks increases linearly with depth and is independent of material. Thus, the brittle shear strength of the lithosphere will increase roughly linearly throughout the crust and mantle part of the lithosphere. In contrast, the viscous strength of rocks is a strong function of (i) material properties (expressed in the viscosity), (ii) temperature (expressed in the Arrhenius relationship) and (iii) strain rate, but it is *independent* of normal stress and therefore independent of depth. However, as the temperature in the crust and mantle lithosphere increases largely monotonously with depth, the viscous shear strength is likely to decrease with depth. Nevertheless it is important to note that this relationship may be reversed if the thermal structure is reversed. Also, as the material changes with depth, for example at the Moho, different viscous curves become relevant and the lithosphere may become rheologically stratified. Such a rheological stratification of the lithosphere is fundamentally different between continental and oceanic lithosphere.

4.1.1. Rheology of the Continental Lithosphere

Within the *continental crust*, the rheology may well be described by the behavior of quartz. Quartz is reasonably soft mineral phase and its shear strength at typical Moho temperatures of the order of 500°C is negligible. If the entire crust deforms at constant strain rate and is characterized by a linear temperature profile, then the viscous rheology of the crust will be described by an exponentially decreasing shear strength with depth. As the brittle strength increases with depth and the viscous strength decreases with depth, the two curves will cross somewhere in the middle crust in a region termed the ”brittle-ductile” transition. Above this region, the brittle shear stresses are smaller than the viscous stresses and the crust will deform in a brittle fashion, below this depth the viscous stresses are smaller than the brittle stresses and the crust will deform viscously. The brittle ductile transition will support the highest shear stresses anywhere in the crust and it is within this depth range that the highest moment release might be anticipated from intraplate earthquakes. At increased strain rate, the viscous stresses will become larger so that the brittle ductile transition moves downwards and vice versa.

Below the Moho, the viscous curve for quartz is irrelevant and is replaced by a corresponding curve for olivine. Olivine supports substantially higher shear stresses than quartz and the Moho is therefore the region of highest strength in the lithosphere. In total, the continental lithosphere may be characterized by two strength maxima: one around the brittle-ductile transition in the middle crust, the other just below the Moho. Deformation in the crust will be vertically, laterally and temporally quite heterogeneous, depending on the mineralogy, presence or absence of fluids, and episodicity of deformation. Much of the deformation, even in the deeper crust and uppermost mantle, is likely to be focused in shear zones, rather than distributed throughout the entire rock mass. Such zones are also likely to act as conduits for any fluid rising (continually or, more likely, episodically) from greater depth. While dry rocks in the deepest crust show strengths that are comparable to mantle rocks, wet crustal rocks may be significantly weaker, resulting in decoupling of the crust from the mantle lithosphere. Thus, the simple curves shown in Figure 5 have been refined by a large range of authors to include more layers within the crust, elastic regimes or a depth dependence of strain rate.

Figure 5. The Brace - Goetze model for the rheology of the lithosphere. ? *d* is the differential stress, and *z* depth. The two diagrams at left show the model for the continental lithosphere and those at right for the oceanic lithosphere. For each model, the left diagram shows the relevant stress curves and the right hand diagram a strength profile. Note that the integrated area under the strength profile for the oceanic lithosphere is larger than that underneath the strength profile for the continental lithosphere. Oceanic lithosphere is therefore stronger.

4.1.1. Rheology of the Oceanic Lithosphere

Oceanic lithosphere behaves rheologically equivalent to continental lithosphere, but there are some important differences. These are largely due to the absence of a crust of significant thickness. A mere 5 to 7 km of oceanic crust is produced at the mid oceanic ridges and the subsequent evolution of oceanic lithosphere is characterized by cooling and ”freezing” of mantle lithosphere to the base of the thin oceanic crust. Thus, the oceanic lithosphere is characterized by a single strength maximum as shown in Figure 5. Note that the integrated region below the strength profile is just as large, if not larger than that for the continental lithosphere at the same strain rate. Thus, oceanic lithosphere is actually stronger than continental lithosphere !

. The article *Rheology of Rocks* by Brian Evans (Massachusetts Institute of Technology, USA) covers many aspects of the study of flow and deformation of different rocks from the mantle as well as the oceanic and continental crust.

4.2. Gravity and Isostacy

Gravity refers to forces exerted on an element of mass at the surface of the earth, It is measured as an acceleration due to gravity. The gravitational acceleration measured on earth has two components:

Gravitational attraction of the mass of the Earth.
Centrifugal forces due to the rotation of earth.

Due to the earth's rotation, topography and lateral variations in the vertical distribution of density the acceleration of gravity the gravitational acceleration varies with the geographical location. A surface of constant gravitational potential is called the geoid. The deviation of this surface from a rotational ellipsoid are called geoid anomalies and are shown on Figure 3. Gravity anomalies measured on the surface of the earth and corrected for terrain are called Bouger anomalies (after Pierre Bouguer, French mathematician, 1698-1758). Mass anomalies in the crust and the mantle that extend over distances greater than a few hundred kilometers are isostatically compensated.

Isostasy is a term derived from the Greek words "iso" and "stasis" meaning "equal standing". The concept of isostasy relates the vertical distribution of mass in the lithosphere with surface elevation. Within this concept, the lithosphere is considered to be floating on the underlying relatively weak asthenosphere. Isostasy is described by a vertical stress balance. This stress balance states that at some given depth, called the isostatic compensation depth, the vertical stresses are all equal. Two different models for compensation have been developed: (i) The Pratt model, named after John Henry Pratt (mathematician and Archdeacon of Calcutta, 1809-1871), proposes that the average density of crustal rocks is lower where the elevation is greatest. (ii) The Airy model, named after George Biddel Airy (Professor of Astronomy at Cambridge University, 1801-1892), proposed that the higher topography is compensated by the presence of crustal root zones. Both hypotheses adequately describe the observed phenomenon although the physical reality probably lies between the two models. For example, most recent orogenic belts tend to have deep crustal roots, but some areas of marked elevations are characterized by hotter, and therefore lower density, underlying mantle rocks than normal.

5. Outlook and Perspectives

Currently, most scientists accept the theory of plate tectonics as the explanation for the movement of plates on the earth's surface. However, when modern plate tectonics was first developed in the 1960s and confirmed by many geodynamic and geophysical observations, less than 0.0001% of the deep ocean had been explored and less than 20% of the land area had been mapped in meaningful detail. Even by the mid-1990s, only about 3 to 5% of the deep ocean basins had been explored in any kind of detail, and not much more than 25 to 30% of the land area could be said to be truly known. Scientific understanding of the earth's surface features is clearly still in its infancy, to say nothing of the earth's interior. Some scientists even think that plate tectonics was a premature generalization of still very inadequate data on the structure of the ocean floor, and has proven to be far removed from geological reality. Nevertheless, many observations have now been explained with geodynamic concepts. If one were to name two unresolved geodynamic problems with the widest reaching implications to tectonic processes within our current understanding of plate tectonics, we might argue that they are:

The driving forces of plate tectonics
The magnitude of differential stress in the lithosphere.

With respect to the driving forces of plate tectonics, there is a debate between one school arguing that plate tectonic motion is exclusively governed by the potential energy of the mid ocean ridges and other potential energy contrast within the lithosphere. This argument rests largely on the well-known potential energy of the mid-oceanic ridges and the reasonably well-known potential energy contrast within the continental lithosphere. These are large enough to drive plate motions. A problem with this model is that the potential energy of mid ocean ridges is insufficient to built the potential energy contrasts observed in the continents. Thus, the less-well known force balances in subduction zones and other processes (for example the focusing of ridge torques) have been considered as important for mountain building. An alternative school argues that the geometry of mid ocean ridges in the past and at present does not correspond with the distribution and direction of most large-scale orogens on our planet. Thus, so it is argued, it must be convection in the mantle and friction at the base of the lithosphere that drives plate tectonics. However, in the past decade, this debate has largely swung to the school arguing for potential energy contrast as the main driving force of plate tectonics.

The discussion of the magnitude of differential stress in the lithosphere is directly related to the discussion of tectonic overpressure that has begun in the 1960's. In this debate, one school argues that experiments show that rocks have insufficient strength to support differential stresses beyond some tens of MPa and that therefore, differential stress has no influence on geobarometry, on frictional heating of orogens or other observable geological parameters. In contrast, there is a strong argument for high shear strength of rocks. This argument lies in a large scale observation: The fact that topographic differences up to 5 km are supported in mountain ranges of our planet gives clear evidence that the mean shear strength of the lithosphere is of the order of 50 MPa. Thus, if the uppermost and lowest parts of the lithosphere are substantially softer than this value, then there must be pats that are substantially stronger.

5.1. Alternative Concepts ?

The theory of plate tectonics is now so widely accepted that it may come as a surprise to discover that some scientists remain unconvinced. Alternative models favored by various scientists include a "rapidly" expanding earth, a "slowly" expanding earth, a contracting earth, and a model called "surge tectonics". This alternative surge theory indicate that expansions and contractions, due to earth oscillations, may supply the force to drive sea floor spreading and other aspects of tectonics. All major features of the earth, including those beneath the sea, are underlain by more or less fluid igneous rocks, which tend to flow parallel to these features. These channels are inter-connected, forming a worldwide hydraulic network. This hypothesis explains a remarkable number of observed earth phenomena. Another extension of the surge theory suggests a link between tectonics and climate incorporating stream flow (geostreams) and vortex formation within the earth’s mantle as part of the driving force. In addition, it is thought that the earth’s internal movement is analogous to ocean and atmospheric models and influenced by similar Coriolis and inertial forces. A link between tectonics and climate is suggested by micro-gravity forcing of atmospheric pressure.

The critics of the theory of plate tectonics comprise several arguments: Probably the most significant is the claim that granitic material is found in the ocean floor, for example not far from the mid-ocean rift system in the Atlantic. If these observations are valid, this would seem to be fatal to the standard interpretation of sea-floor spreading. However, it has to be proven that the continental rocks reported from along the mid-ocean rift system were *in situ*, rather than transported from continents. The strongest claim seems to be Bald Mountain near the Azores, which is close to the ridge. The presence of autochthonous continental material here would seem to require a change in the standard interpretation of plate tectonics.

Questions are also raised about magnetic properties of ocean-floor rocks. The nature of magnetism seems related to other phenomena. For example, normal magnetic polarity seems associated with high heat flow, hot-spot volcanism, fast sea-floor spreading, and rapid rates of subsidence of cratonic basins. Volcanic quiescence seems correlated with reverse magnetic fields, yet while most of the Hawaiian-Emperor chain was formed during normal polarity, reverse polarity is considered to have occupied an equal amount of time during their formation.

Another criticism is that intraplate activity is not explained by plate tectonics, a fact that is already well-known and proven by geophysical and geodetic methods. Examples include: formation of intracratonic basins and mid-continental earthquakes, such as the New Madrid earthquake, which occurred in 1811 and 1812 near New Madrid (Missouri). These seismic events had a magnitude of 8.0 or higher on the Richter Scale and are among the great earthquakes of known history, affecting the topography more than any other earthquake on the North American continent. Furthermore, the mechanism driving plate movement has never been satisfactorily determined. The criticism here is that ridge push seems to be more important than slab pull. For example, subduction pull, where old, cold dense oceanic lithosphere sinks under its own weight pulling the plate after it, cannot explain how continental collision of India with Asia could result in formation of the Himalayas. Slab pull would also fail to explain the driving force, which still pushes India below the Himalayas at a rate of some cm a^{-1}.

Scientists arguing against the principles of plate tectonics are undoubtedly outside the mainstream of current thinking, and some of the arguments presented above are not persuasive. However, many details of geodynamic processes are still controversial or unknown, reminding us that theories may appear to be well established, yet may have significant shortcomings. New ideas like ”platonics”, where the mosaic of plates is considered to be a self-organizing network of plates and force chains, which are readily reorganized by stress changes and have surficial explanations rather than mirrors of the planform of mantle convection, will certainly help to continue to develop the idea of plate tectonics resulting in a better understanding of its geodynamic cause.

5.2. The Future of Geodynamics

Since the advent of plate tectonics about 50 years ago, enormous advances have been made in geodynamics and tectonics by applying extremely simple analytical models to tectonic problems of the largest scale. The most spectacular of all is undoubtedly the age-depth heat flow relationship of the oceanic lithosphere discussed in section 3.1.1. However, models like the Brace Goetze lithosphere discussed in section 4.1., the simple models for Barrovian and Buchan style metamorphism and countless more follow a close second. Most of these models have in common that very simple solutions of the differential equations governing force and energy balances have been used to explain single (if omnipresent) observations.

However, in the past decade, it has become more and more common to use large data sets containing global sets of observations as the basic observational data on which models are based. We believe that the future will be more and more characterized by the use of such data sets and the use of numerical rather than analytical models to explain them.

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Glossary

Advection: The transfer of heat energy by means of mass motions through a medium described by a velocity field.

Asthenosphere: The relatively plastic layer of the upper mantle of the Earth on which the tectonic plates of the lithosphere move. The asthenosphere is approximately 200 km thick and ~ 1400 °C, but not molten. Here the mantle deforms by plastic flow in response to applied stresses above 100 MPa.

Brittle: Deformation due to fracture. Assuming a geothermal gradient of ~30°C km^{-1}, brittle failure is generally the dominant deformation process within the upper 10 km of the Earth's crust.

Cohesion: Ability of particles to stick together without dependence on interparticle friction (e.g., compaction or a binding substance).

Conduction: Heat energy transfer directly from atom to atom representing the flow of energy along a temperature gradient.

Core: Innermost layer of the Earth. The core makes up nearly 3500 km of the earth's radius of 6370 km and its outer part is supposed to be partly liquid. The inner core is solid and 1220 thick and consists mainly of iron and nickel.

Crust: The thin, outermost shell of the Earth that is typically 5 km to 75 km thick. The continental crust comprises rocks similar in composition to granite and basalt (i.e., quartz, feldspar, biotite, amphibole and pyroxene) whereas the composition of oceanic crust is basaltic (pyroxene and feldspar). The crust overlies the more dense rock of the mantle.

Deformation: Change in shape and orientation of volumes of rocks from an initial to its final state. See strain.

Density: Mass per unit of volume. Density is typically reported in kg m^{-3}.

Dislocation: A linear crystalline defect around which there is atomic misalignment; plastic deformation corresponds to the motion of dislocations in response to an applied shear stress; edge, screw, and mixed dislocations are possible.

Ductile: General term for permanent deformation that is not by brittle fracture. Typical deformation mechanism in the ductile regime are viscous deformation and plastic deformation.

Elastic: The deformation that can be recovered when an applied stress has been removed. When the elastic limit of a material has been exceeded, non-recoverable, permanent deformation occurs.

Eigenvalue: An eigenvalue of an n by n matrix A is a scalar c such that A x = c x holds for some nonzero vector x (i.e., Eigenvector).

Eigenvector: An eigenvector of a square matrix A is a nonzero vector x such that A x = c x holds for some scalar c (i.e., Eigenvalue).

Exhumation: Displacement of rocks with respect to the surface. The rate of exhumation is the rate of erosion or the rate of removal of overburden by tectonic processes. See uplift.

Fault: A discontinuity in rocks across which there is observable displacement. Depending on the relative direction of displacement between the rocks, or fault blocks, normal-, thrust- (reverse) or strike-slip faults are distinguished.

Fold: A wave-like geologic structure that forms when rocks deform by shortening and bending instead of breaking under compressional stress.

Geochronology: The determination of the absolute age of rocks, minerals and fossils, in years before the present. The measurement of the decay of radioactive isotopes, like uranium, potassium, rubidium, thorium and carbon, has allowed geologists to more precisely determine the age of rock formations. Additionally tree rings and seasonal sedimentary deposits (varves) can be counted to determine absolute age. Although the term implies otherwise, "absolute" ages typically have some amount of potential error and are inexact.

Geodesy: Geodesy is the science of mathematically determining the size and shape of the earth and the nature of the earth's gravity field.

Geoid: Equipotential surface of equal gravity magnitude of the earth, which is always perpendicular to the local gravity vector.

Geomorphology: The systematic description, analysis, and understanding of landscapes and the processes that change them (e.g., incision of rivers, landslides, erosion).

Geothermal gradient: The rate of increase in temperature per unit depth in the earth. Although the geothermal gradient may vary significantly from place to place and is strongly dependent on exhumation rates, it averages within the stable continental crust 25 to 30 °C km^{-1}.

Global Positioning System (GPS): A system of numerous earth-orbiting satellites that can be used to determine the location (latitude, longitude and elevation) of a receiver or station on the earth within about 2 m. Fixed receivers on earth can be used to increase the accuracy and to determine the relative motions of fault blocks and lithospheric plates.

Gravity: The earth's gravitational field, or the attractive force produced by the mass of the earth. Variations in the gravitational field can be used to map changes in the density of formations in the Earth.

Hot spot: A volcanic center (~ 100 - 200 km) persistent for at least a few tens of Myr, that is thought to be the surface expression of a persistent rising plume of hot mantle material. Hot spots are not linked to arcs and may not be associated with mid ocean rift systems.

Isostacy: The buoyant condition of the earth's crust floating in the asthenosphere.

Isotherm: A line (or plane) of equal or constant temperature.

Lithosphere: The stiff outer layer of the earth that includes the crust and uppermost mantle. It is made up of several tectonic plates that move around on the softer asthenosphere. The lithosphere of the oceans tends to be thinner (~ 50 km) and more dense than that of the continents (~120km) because of isostasy.

Ma (a): Million years (years) referring to an age. See Myr.

Mantle: The intermediate layer of the earth beneath the crust that is about 2900 km thick and overlies the core of the earth. The mantle consists of dense rocks composed of the minerals pyroxene and olivine. The crust, mantle and core of the earth are distinguished from the lithosphere and asthenosphere on the basis of their composition and not their mechanical behavior.

Mid ocean rift systems (midoceanic ridges): The mountainous, linear axis of topographic high on the ocean floor along which rifting (spreading) occurs and new oceanic crust forms by solidification of basaltic magma derived from the mantle (midocean ridge basalt, or MORB). The presence of symmetrically spreading plate boundaries, the occurrence of successively older crust outward from the ridges and the lack of oceanic crust older than approximately 200 Ma support the theory of plate tectonics and the recycling of oceanic crust through the process of subduction.

Myr (yr): Million years (years) referring to a time span. see Ma.

Neotectonic: Active deformation of the earth's crust including tectonics processes of the last few Myr that have influenced the active deformation pattern.

Orogen: Mountain belt that formed as a result of plate tectonic activity in which lithospheric plates collide often associated with subduction and igneous activity. Faults and folds, crustal thickening, exhumation of metamorphic rocks and surface uplift are typical geological structures and processes of orogens.

Plastic: A ductile deformation mechanisms that occurs after a certain threshold stress has been reached, but is thereafter independent of strain or strain rate. Materials like sand fractured rocks deform plastically on a large scale.

Plate: A segment of the lithosphere, which has little volcanic and seismic activity but is bounded by almost continuous belts (i.e., plate margins) of earthquakes associated with subduction, rifting or large faults. Although there is still an increasing number of newly discovered small plates, most Earth scientists agree that there are seven large major plates (i.e., African, Antarctic, Eurasian, Indo-Australian, North American, Pacific and South American).

Plane strain: Three-dimensional flow for which there is no length change along the intermediate principal stretching direction.

Pure shear: Special type of plane strain flow for which the eigenvectors of the deformation tensor remain parallel to the principal stress axis (coaxial deformation).

Rheology: Generally, the study of how matter deforms, including its elastic, plastic and viscous flow. In geology, rheology is particularly important in studies of moving rock analogues (e.g., ice, plasticene, sand), as well as in studies of deforming rocks.

Ridge: see mid ocean rift systems.

Ridge push: Gravity acting on the topography of mid-ocean rift systems tries to spread the ridge outward. This pushes the rest of the oceanic plate away from the ridge.

Rifting: Process by which the lithosphere is pulled apart, associated with the creation of normal faults, surface uplift and subsidence.

Seismic: Pertaining to waves of elastic energy, such as that transmitted by P-waves and S-waves, in the frequency range of approximately 1 to 100 Hz. Seismic energy is studied by scientists to interpret the composition, fluid content, extent and geometry of rocks in the subsurface.

Seismology: The study of seismic or elastic waves, such as from earthquakes, explosions or other causes. Interpretation of the structure and composition of the earth from artificially created seismic waves is a chief concern of seismologists exploring for hydrocarbons and other resources.

Simple shear: Special type of plane strain flow for which a plane exists (i.e., flow plane), which is not deforming. This plane lies at 45° to the principal stress axes.

Slab pull: Old, cold dense oceanic lithosphere sinks under its own weight. As the lithosphere is strong, the negative buoyancy of the slab pulls the plate after it.

Specific heat: A material property that indicates the amount of energy a body stores for each degree increase in temperature, on a per unit mass basis. Its units are J kg-K^{-1}.

Strain: Tensor quantity describing the permanent change in shape of rocks and other solid bodies. A change in shape, such as folding, faulting, fracturing are common examples of strain seen in rocks. Strain is commonly represented by an ellipsoid comparing with an unstrained sphere. Strain is a more restricted term than deformation, which also includes rotational and translational components.

Stress: Tensor quantity describing orientation and magnitude of force vectors acting on planes of any specific orientation at a specific point in a rock volume.

Subduction: A plate tectonic process in which one lithospheric plate descends beneath another into the asthenosphere during a collision at a convergent plate margin. Because of the relatively higher density of oceanic lithosphere, it will typically descend beneath the lighter continental lithosphere during a collision. In a collision of plates of continental lithosphere, the density of the two plates is so similar that neither tends to be subducted and orogens form.

Tensor: Generalization of vectors and matrices. A tensor of rank 0 is a scalar, of rank 1 a vector, rank 2 is a matrix, rank 3 a three-dimensional rectangular array and rank k a k-dimensional rectangular array.

Thermal diffusivity: A material property that describes the rate at which heat diffuses through a body. It is a function of the body's thermal conductivity and its specific heat. Its units are $m^2\ s^{-1}$.

Trench: Deep, linear zones that form where an oceanic plate sinks beneath another plate (subduction zone).

Triple junction: The point or region where three plate boundaries meet at roughly 120° angles on a map. They form during the initial stages of continental rifting. As rifting continues, two of the three "arms" of the triple junction continue to widen, perhaps creating a new ocean basin between the separating continental chunks, while the third one "fails" (i.e., aulacogen).

Uplift: Displacement component in the direction opposite to the gravity vector. Surface uplift is the displacement of the earth's surface with respect to the geoid. Uplift of rocks is the displacement of rocks with respect to the geoid. See exhumation.

Viscosity: A fluid property that relates the magnitude of shear stresses to the strain rate, or more simply, to the spatial rate of change in the fluid velocity field.

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Biographical Sketches

Bernhard Grasemann is an associate Professor for structural geology at the University of Wien in Austria. He special expertise is the coupling of fieldwork with numerical modeling. He is currently working in the Himalayas, the Aegean region and the Alps and has widely published on theoretical questions in structural geology and tectonics.

Kurt Stüwe is an associate Professor for geodynamics at the University of Graz in Austria. He has been educated at the Universities of Melbourne and previously held positions at Monash University and Adelaide University in Australia. His specialty is interdisciplinary research between petrology and structural geology, and he is particularly interested in numerical modeling. He has worked in all continents and has written two textbooks on geodynamics.

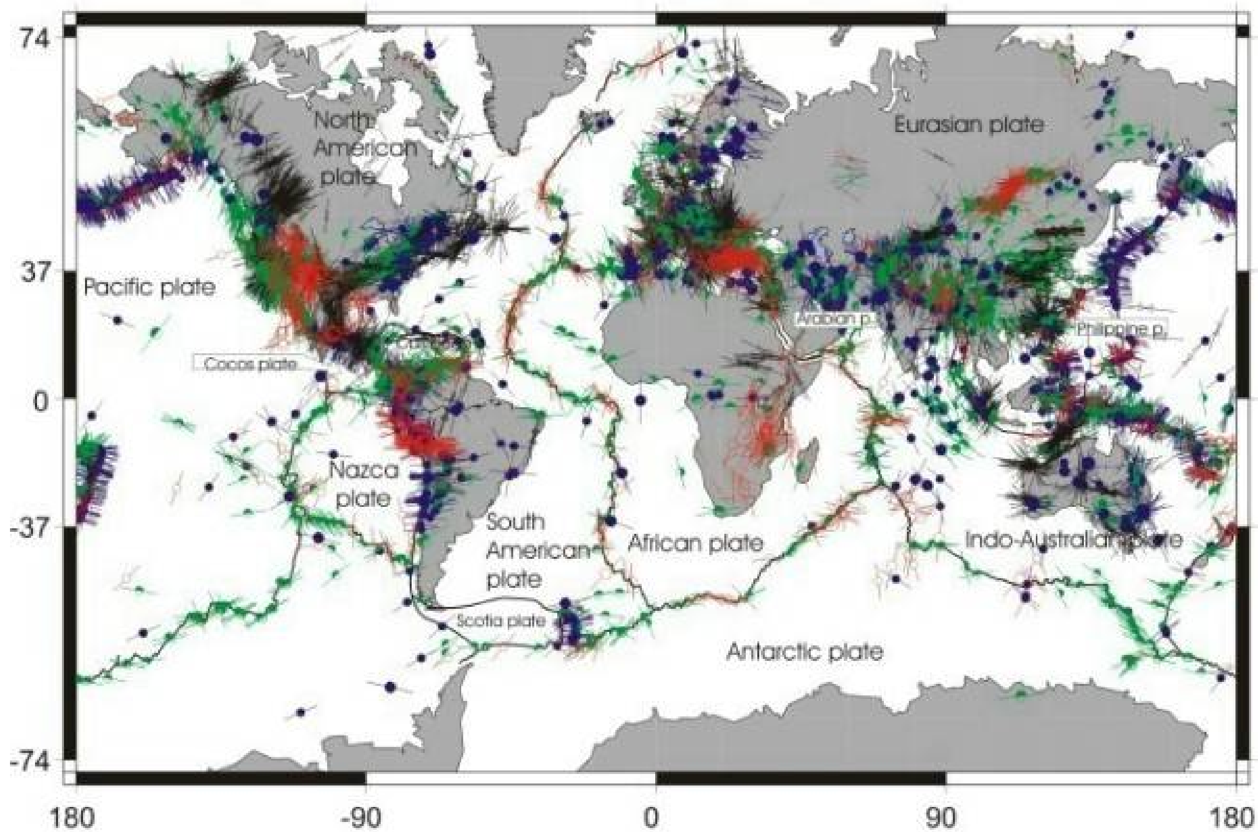

Figure 1: The intraplate stress field of the world superimposed on a rough plate tectonic subdivision of earth. Different symbols indicate different methods of stress determination including earthquake focal mechanisms, borehole breakouts and geological indicators. Different shadings indicate different deformation regimes: darkest are thrust faults, medium grey are normal faults and light shading are strike slip faults. This map was created using the CASMO facility on the world stress map project home page. The map is also modified after Stüwe (2002).

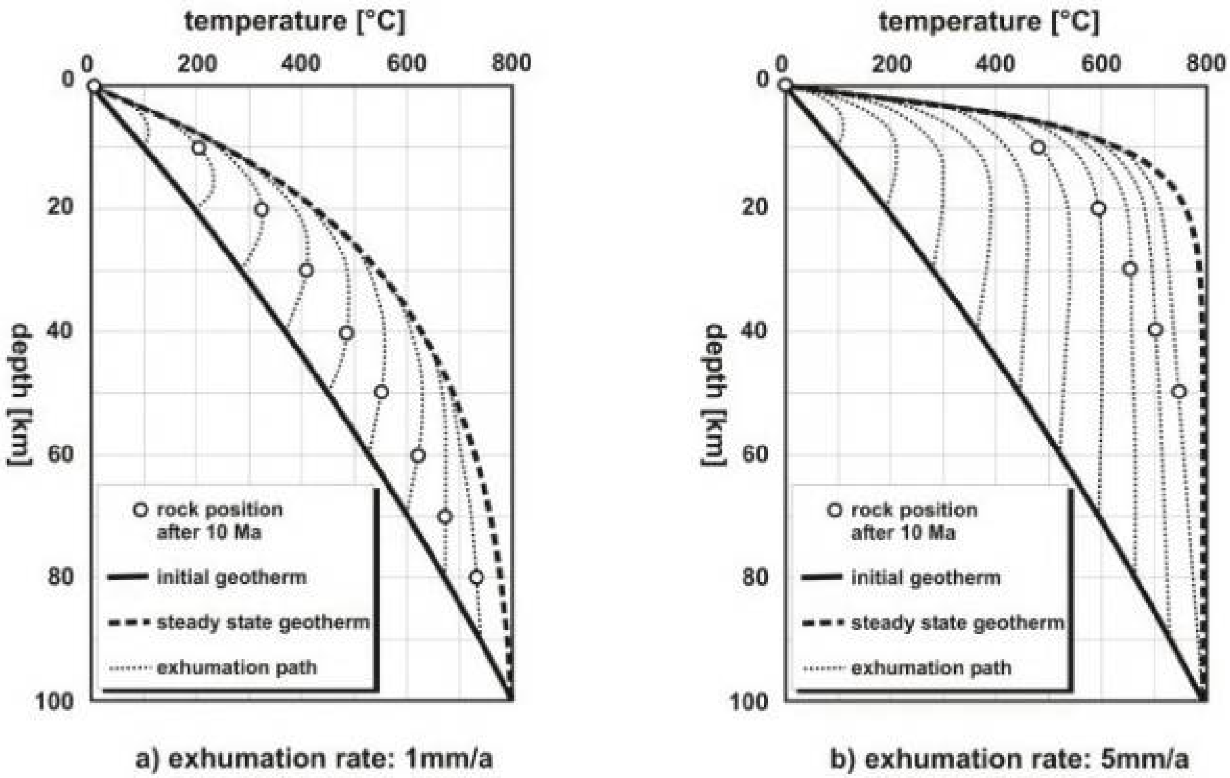

Figure 2: Modelled exhumation (temperature-depth) path of rocks exhuming from 10-100 km depth at (a) 1 mm/a and (b) 5 mm/a. Note that higher exhumation rates favour isothermal decompression of rocks (modified after Grasemann et al. 1998).

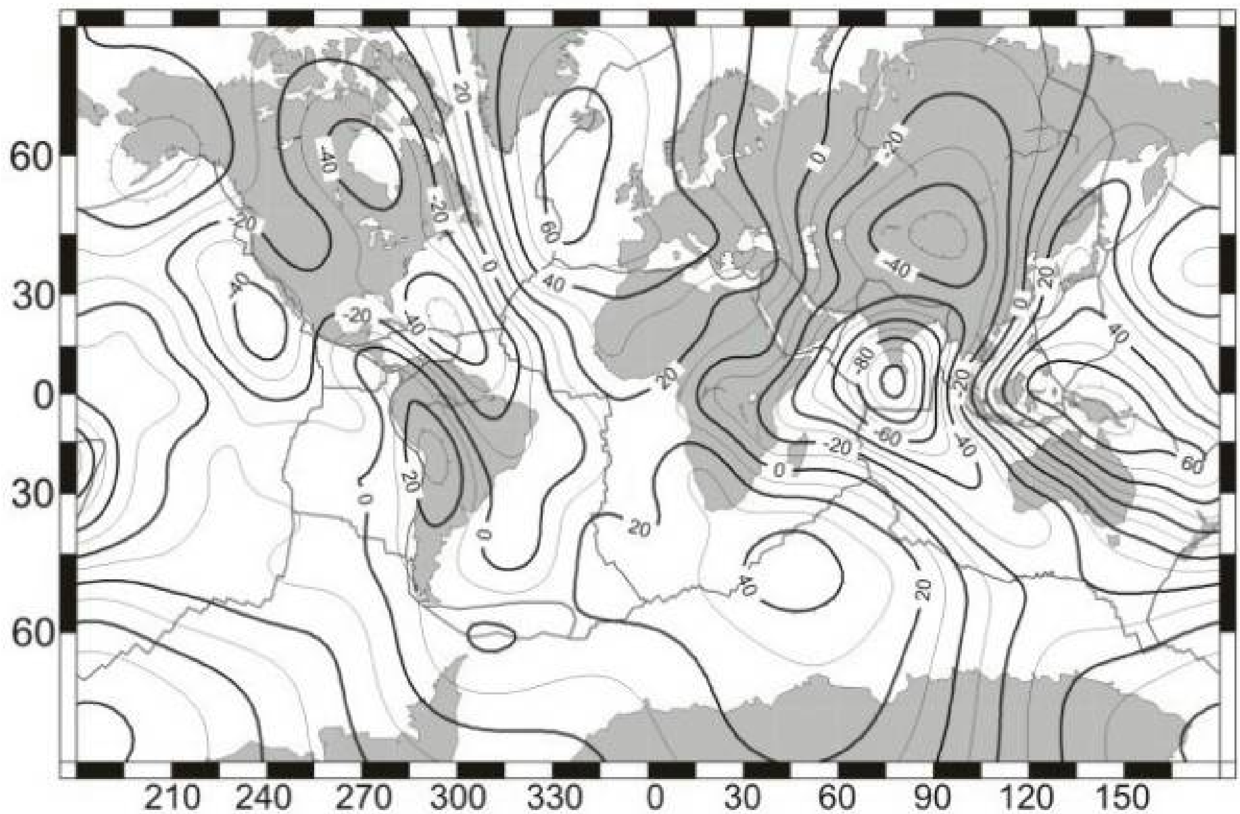

Figure 3. Geoid map of the world. Modified after Stüwe (2002).

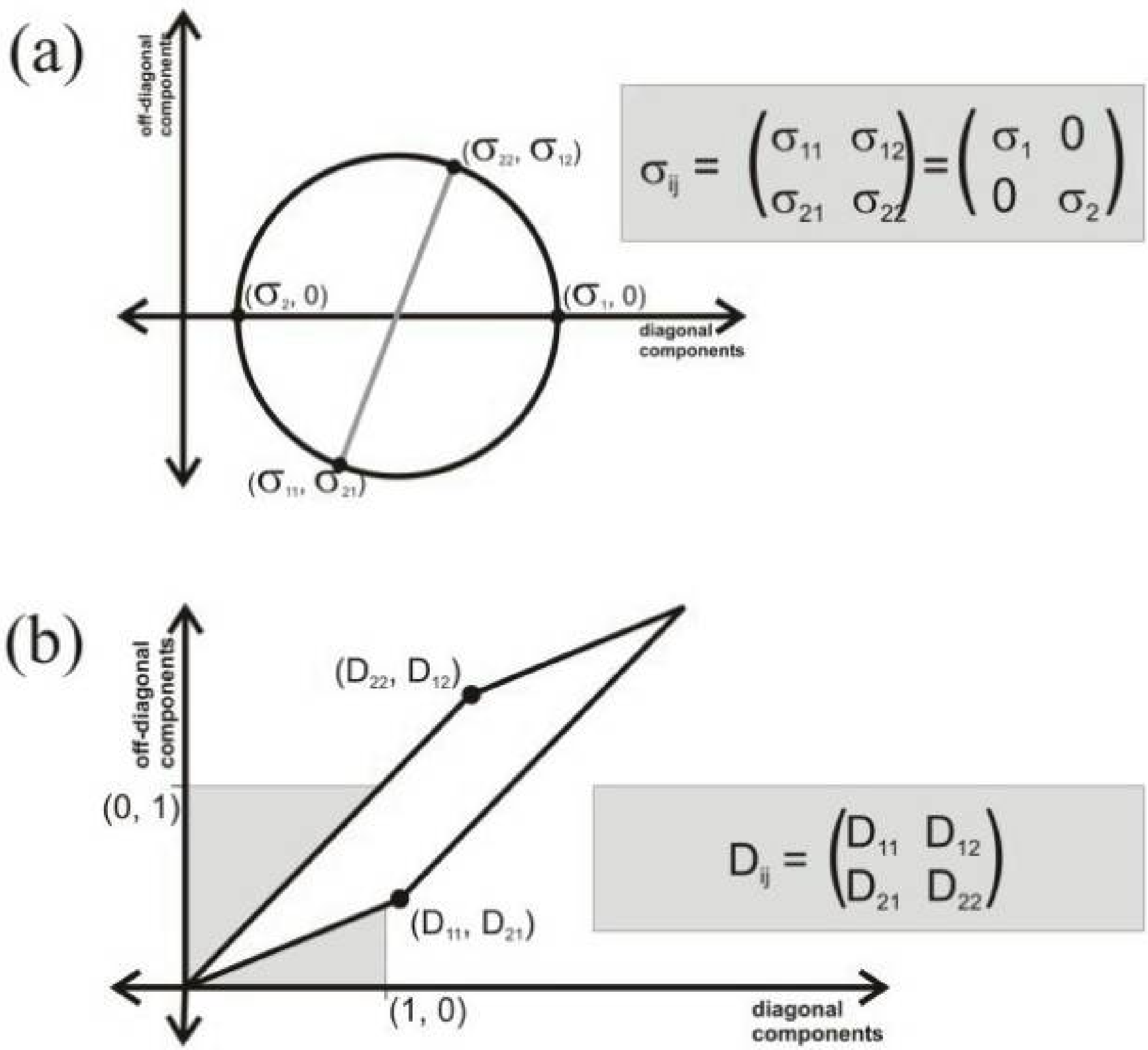

Figure 4. (a) Mohr Circle construction of a symmetric tensor S. (b) Deformation tensor D relating two opposite corners of a unit square with the same corners of its deformed equivalent.

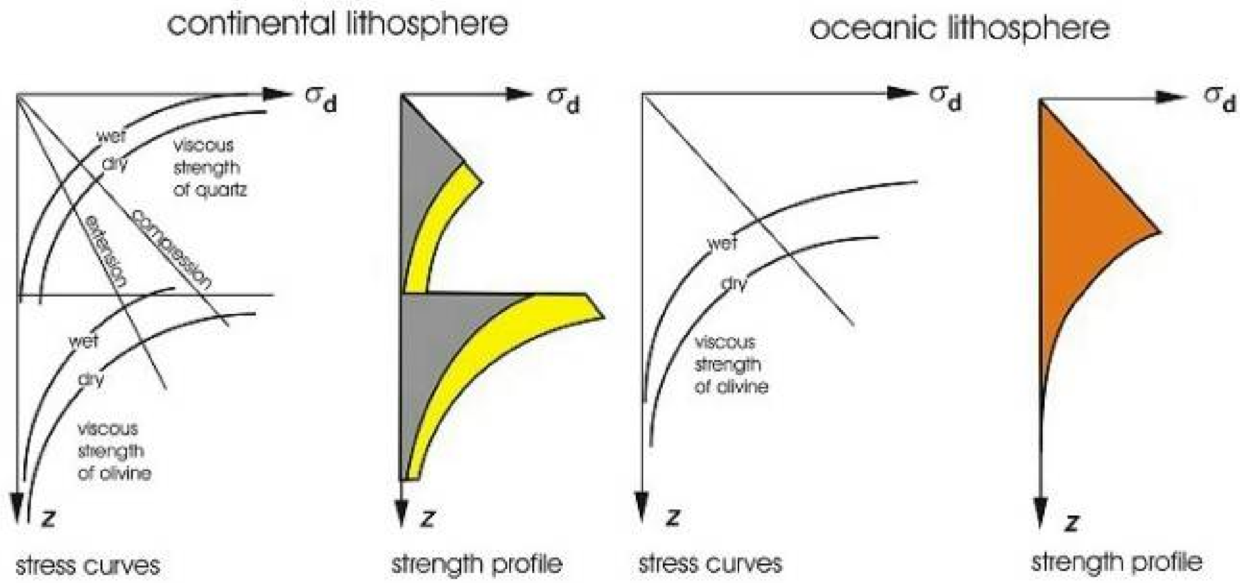

Figure 5. The Brace - Goetze model for the rheology of the lithosphere. sd is the differential stress, and z depth. The two diagrams at left show the model for the continental lithosphere and those at right for the oceanic lithosphere. For each model, the left diagram shows the relevant stress curves and the right hand diagram a strength profile. Note that the integrated area under the strength profile for the oceanic lithosphere is larger than that underneath the strength profile for the continental lithosphere. Oceanic lithosphere is therefore stronger.

