It is widely accepted that the Earth's interior is composed of several layers: the crust, the mantle and the core. Since the crust is readily accessible, scientists have been able to perform hands-on experiments to determine its composition; studies on the more distant mantle and core have more limited opportunities samples, so scientists also rely on analyses of seismic waves and gravity, as well as magnetic studies. Scientists can analyze the Earth's crust directly, but they rely on seismic and magnetic analyses to investigate the Earth's interior. Where the crust has been disturbed, it is easy to see layers of different materials that have settled and compacted. Scientists recognize patterns in these rocks and sediment, and they can evaluate the composition of rocks and other samples taken from different depths of the Earth during routine excavation and geologic studies in the lab. The United States Geological Survey Core Research Center has spent the past 40 years amassing a rock core and cuttings repository and making these samples available for study. Rock cores, which are cylindrical sections brought to the surface, and cuttings (sand-like particles) are kept for potential re-analysis as improving technology allows for more in-depth study. In addition to visual and chemical analyses, scientists also try to simulate conditions deep under the Earth's crust by heating and squeezing samples to see how they behave under those conditions. More information about the Earth's composition comes from studying meteorites, which provide information about the likely origin of our solar system. It is impossible to drill to the center of the earth, so scientists rely on indirect observations of matter lying below the surface through use of seismic waves and their knowledge of how these waves travel during and after an earthquake. The speed of seismic waves is affected by the properties of the material the waves pass through; the stiffness of material affects the speed of these waves. Measuring the time it takes for certain waves to get to a seismometer after an earthquake can indicate specific properties of the materials that the waves encountered. Where a wave encounters a layer with a different composition, it will change direction and/or velocity. There are two types of seismic waves: P-waves, or pressure waves, which go through both liquids and solids, and S-waves, or shear waves which go through solids but not liquids. P waves are the faster of the two, and the gap between them provides an estimate of the distance to the earthquake. Seismic studies from 1906 indicate that the outer core is liquid and the inner core is solid. Yavorski, Kimberly. "How Do Scientists Know The Structure Of The Earth's Interior?" sciencing.com, . 23 April 2018. APA Yavorski, Kimberly. (2018, April 23). How Do Scientists Know The Structure Of The Earth's Interior?. sciencing.com. Retrieved from Chicago Yavorski, Kimberly. How Do Scientists Know The Structure Of The Earth's Interior? last modified March 24, 2022. Three major layers that make up the interior structure of the Earth are crust, mantle, and core each having a distinct chemical composition and physical state. The interior of the Earth can be divided into several layers based on the chemical composition and mechanical stability. Based on these characteristics, Earth’s internal structures are divided into crust, lithosphere, asthenosphere, mantle, outer core and inner core. Except for the liquid outer core, all the other layers are solid, including the semi-molten asthenosphere. The difference in the chemical and physical nature of the Earth’s layers is due to differential temperature, pressure and density which have been shaped during the evolution of these layers. The internal structures of Earth have been proposed based on direct and indirect sources such as seismic waves which behave differently in the Earth’s layers depending on the composition and physical nature of these layers. Sources for the Interior of the Earth The materials of the earth’s surface are different from the interior. Therefore, to study the Earth’s internal structure, direct and indirect sources are taken into consideration. Direct Sources: Mining: The most readily available earth material is its minerals, which can be obtained through mining. Mines can be very deep also, for example, Gold mines in South Africa are as deep as 3 – 4 km. Drilling Projects: There are many projects to penetrate deeper depths to explore the conditions in the crustal portions such as the “Deep Ocean Drilling Project” and “Integrated Ocean Drilling Project”. Volcanic eruption: During a volcanic eruption, molten material (magma) explodes onto the earth’s surface and is then available for laboratory investigation. However, it is difficult to determine the depth of the magma’s source. Indirect Sources: Temperature, Pressure, and Density: Knowing the total thickness of the earth, scientists have estimated the values of temperature, pressure, and the density of materials at different depths. Seismic Activity: The shadow zones of P- and S- waves help understand the different densities of the Earth’s layers. For example, the S-waves cannot be transmitted through fluids. The sudden “shadow” where s-waves disappear indicates that the Earth had a liquid outer core. Different discontinuities within the Earth’s interiors are based on the propagation and velocities of seismic waves through the layers of the Earth. Gravitation: The gravitational force differs at different latitudes on the Earth. Gravity anomalies provide information on the mass distribution of materials in the Earth. Magnetic surveys: These also offer information regarding the distribution of magnetic materials in the crustal portion, as well as the distribution of materials in this region. Meteors: The material and the structure observed in the meteors are similar to that of the Earth. Hence, this becomes a source of information about the interior of the Earth. Structure of the Earth Earth is made up of several layers. According to the mechanical properties, Earth’s layers are the Lithosphere, Asthenosphere, Lower mantle (also known as mesospheric mantle), Outer core, and Inner core. Chemically, the layers are the Crust, Mantle, and Core. Evolution of the Earth’s Layered Structure When Earth was formed about 4.5 billion years ago, it was a uniform ball of hot rock. Radioactive decay and leftover heat from planetary formation caused the ball to get even hotter. Iron Catastrophe: Eventually, after about 500 million years, Earth’s temperature heated to the melting point of iron—about 1,538° Celsius. This key stage in Earth’s history is known as the Iron Catastrophe. The iron catastrophe allowed more rapid movement of Earth’s molten, rocky material. Formation of layers:The early core was formed when droplets of iron, nickel, and other heavy metals gravitated to the Earth’s center to form the core of the new planet. Relatively buoyant materials, such as silicates, and water, stayed close to the planet’s exterior. This process is known as planetary differentiation. The molten material that surrounded the core was the early mantle. Materials that initially stayed in their liquid phase during this process, called “incompatible elements,” ultimately became Earth’s brittle crust. The crusts are still evolving due to plate tectonics. Solidification of Mantle: Water held within minerals erupted as lava, a process known as “outgassing.” As more water was expelled, the mantle solidified. Thermal and Physical State of Earth’s Interior Temperature, pressure, and density are responsible for the generation of the present state of affairs. Temperature: In deep wells and mines, the temperature rises as the depth increases.This evidence along with molten lava erupting from the earth’s interior, supports that temperature increases towards the center of the earth. The pace of increase is not linear. Within the upper 100 kilometers, the temperature gradient is between 15° and 30°C per kilometer. It then decreases substantially through the mantle, increases more swiftly at the base of the mantle, and finally gradually increases through the core. The temperature is around 1000°C at the crust’s base, 3500°C at the mantle’s base, and 5,000°C at the Earth’s center. The disintegration of radioactive elements and chemical reactions occurring under high pressure may be the reason for the earth’s extremely high temperature. Pressure: Due to the enormous weight of the overlying rocks, there is also an increase in pressure from the surface of the earth towards its core. As a result, the pressure is extremely high in deeper areas. The pressure near the center is considered to be 3 to 4 million times thepressure of the atmosphere at sea level. Density:On average, the density of the Earth’s interior is 5.5 g/cm3. Due to the increase in pressure and presence of heavier materials towards the earth’s center, the density of the earth’s layers also increases. Crust The crust is the uppermost layer of the Earth. It possesses just 1% of Earth’s mass but contains almost all known life in the universe. The crust, created by the dynamic geologic forces, continues to be shaped by the planet’s movement and energy. Plate tectonics is responsible for the formation and destruction of crustal materials. Composition: It is made up of solid rocks and minerals. Earth’s crust is composed of igneous, metamorphic, and sedimentary rocks. Igneous rocks such as granite and basalt, which form when magma cools, are the most prevalent types of rocks in the crust. Density: less than 2.7 g/cm3. Types: The crust is subdivided into two types – oceanic and continental crusts. The oceanic crust is found under oceans. The continental crust floats higher on the mantle because it is less dense than the oceanic crust. Discontinuity: The transition zone between oceanic and continental crust is known as Conrad discontinuity. Difference between Oceanic and Continental Crust Features Oceanic Crust Continental Crust Composition – Mafic and ultramafic intrusive igneous. – Rocks of the oceanic crust referred to as “SiMa”, stand for Silicate and Magnesium, the most abundant minerals in the oceanic crust. – Granitic (felsic) intrusive igneous rocks. – The term “SiAl” refers to rocks of the continental crust that contain silicate and aluminium, the most abundant minerals in the crust. Formation – Formed from magma that rose from the mantle and cooled at the ocean floor. – Formed from the melting of rocks and the accumulation of sediments. Mineralogy – Rich in iron and magnesium – Rich in silicon and oxygen Thickness – Oceanic crust is typically about 5-10 kilometers thick. – Continental crust can be up to 40 kilometers thick. Density – 2.9 grams per cubic centimeter – 2.7 grams per cubic centimeter Mantle It is the 2,900 km thick layer between Earth’s dense, superheated core and its thin outer layer, the crust. Volume: The mantle lies below the crust and is by far the largest layer making up 84% of Earth’s volume and 67% of the Earth’s mass. Density: The density of the mantle is3.9 g/cm3. Composition: The majority of the rocks that make up the mantle of the Earth are silicates. Olivine, garnet, and pyroxene are the common silicates found in Mantle. Magnesium oxide is the other main type of rock that can be found in the mantle. Other elements include aluminium, iron, calcium, sodium, and potassium. Discontinuity: The Mohorovicic Discontinuity, or Moho, marks the beginning of the mantle. The Moho is defined as the density contrast from less dense crust to the denser mantle and whereseismic wave velocities increase. Layers of Mantle The Earth’s Mantle is divided into mainly two layersi.e., the Upper mantle and the Lower mantle. The discontinuity between the upper mantle and the lower mantle is known as Repetti Discontinuity. Upper Mantle: Despite being mostly solid, the upper mantle’s more malleable regions contribute to tectonic activity. The thickness of the Upper mantle is around 410 Kms. Two parts of the upper mantle are often recognized as distinct regions in Earth’s interior i.e., the lithosphere and the asthenosphere. Transition zone: It is present at the depth from 410 Km to 660 Km. It prevents the large exchanges of material between the lower and upper mantle. The most important aspect of the mantle’s transition zone is its abundance of water. Crystals in the transition zone hold as much water as all the oceans on Earth’s surface. Lower Mantle: The lower mantle extends from about 660 kilometers to about 2,700 kilometers beneath Earth’s surface. Compared to the upper mantle, the lower mantle is hotter and denser. D’’ (D-double prime): It is a shallow region beneath the lower mantle. In some areas, it is a razor-thin boundary with the outer core while in others, it has thick accumulations of iron and silicates. Lithosphere It is the solid, outer part of Earth. It is made up of the crust and the upper mantle, above the asthenosphere. Of all the layers of the Earth, the lithosphere is both the coolest and the most rigid. Thickness: The average thickness of the lithosphere is 100 km and may go up to 300 km below the orogenic mountains. The thickness of the lithosphere is less than 50 km below the oceanic crust. Composition: The lithosphere consists of many different large segments or blocks, called lithospheric plates or tectonic plates. These plates are considered rigid bodies floating horizontally over the asthenosphere, and tectonic deformations typically occur at the plate boundaries as a result of plate interactions with other plates. Asthenosphere The asthenosphere is a hot, soft, mechanically weak, ductile and semi-viscous region and consists of semi-molten rock materials. Properties: The Asthenosphere is part of the upper mantle also known as the Low-Velocity Zone (LVZ) because the velocity of seismic wavesdecreases in this zone. This zone allows the lithospheric plate to float and move over it. Thickness: The average thicknessis between 180 to 220 km. Depth: It lies below the lithosphere at an average depth of 100 km and extends to a depth of 350 to 650 km. Composition: It is composed of peridotite rock, containing mostly the olivine and pyroxene minerals. Core The extremely hot and dense center of our planet is known as the Earth’s core. The Gutenberg discontinuity signals the end of the mantle and the commencement of Earth’s liquid outer core. Volume: The core accounts for 33% of the Earth’s mass and 16 percent of the Earth’s volume. Composition: Unlike the mineral-rich mantle and crust, it is made almost entirely of metal – iron (Fe) and nickel (Ni) hence, sometimes called NiFe layer. Siderophiles, the elements that dissolve in iron (gold, platinum, cobalt, etc) are also found in the core. The core contains 90% of the earth’s sulfur. The other elements speculated to be the parts of the core are oxygen and silicon. Two layers: The core is further divided into two layers – inner core and outer core. The Lehmann discontinuity or the Bullen discontinuity is the boundary separating these regions. Outer Core The 2,200 km thick outer core is composed of liquid iron and nickel, in a molten state. Properties: The hottest part of the core (at the Bullen discontinuity) is as hot as the surface of the sun (around 6,000° Celsius). The liquid metal in the outer core has low viscosity. Earth’s magnetic field: The Earth’s magnetic field is created in the outer core by a self-exciting dynamo process. The magnetic field is generated by electrical currents flowing through slow-moving molten iron. Inner core Earth’s inner core is solid andextends from 5150 Km to 6370 Km, and is mostly composed of iron. Solid core: Despite the temperature of the inner core being more than the melting point of iron, the inner core is solid. This is due to the intense pressure and density of the inner core. Rotation: The inner core rotates eastward but at a little faster rate than the other part of the Earth. Growth: The inner core grows by about a millimeter every year as the Earth is slowly cooling. Consequently, the outer core is solidifying or freezing. Q1. What constitutes the Earth’s structure?+Ans. Three major components make up the Earth's structure - the crust, the mantle and the core. Each layer has a distinct chemical composition, physical state, and potential to affect life on the surface of the Earth.Q2. What are Earth's interior properties?+Ans. The Earth's mantle is made of solid/plastic while the inner core is solid and the outer core is liquid. This is caused by the different layers' varying melting points (the nickel-iron core, the silicate crust, and the mantle), as well as the rise in temperature and pressure with increasing depth.Q3. Why is Earth’s Core hot?+Ans. The primary contributors to heat in the core are the decay of radioactive elements, leftover heat from planetary formation, and heat released as the liquid outer core solidifies near its boundary with the inner core. Additional text to be added here.Tags: interior of the earth sources quest Share — copy and redistribute the material in any medium or format for any purpose, even commercially. Adapt — remix, transform, and build upon the material for any purpose, even commercially. The licensor cannot revoke these freedoms as long as you follow the license terms. Attribution — You must give appropriate credit , provide a link to the license, and indicate if changes were made . You may do so in any reasonable manner, but not in any way that suggests the licensor endorses you or your use. ShareAlike — If you remix, transform, or build upon the material, you must distribute your contributions under the same license as the original. No additional restrictions — You may not apply legal terms or technological measures that legally restrict others from doing anything the license permits. You do not have to comply with the license for elements of the material in the public domain or where your use is permitted by an applicable exception or limitation . No warranties are given. The license may not give you all of the permissions necessary for your intended use. For example, other rights such as publicity, privacy, or moral rights may limit how you use the material. Three centuries ago, the English scientist Isaac Newton calculated, from his studies of planets and the force of gravity, that the average density of the Earth is twice that of surface rocks and therefore that the Earth's interior must be composed of much denser material. Our knowledge of what's inside the Earth has improved immensely since Newton's time, but his estimate of the density remains essentially unchanged. Our current information comes from studies of the paths and characteristics of earthquake waves travelling through the Earth, as well as from laboratory experiments on surface minerals and rocks at high pressure and temperature. Other important data on the Earth's interior come from geological observation of surface rocks and studies of the Earth's motions in the Solar System, its gravity and magnetic fields, and the flow of heat from inside the Earth. The planet Earth is made up of three main shells: the very thin, brittle crust, the mantle, and the core; the mantle and core are each divided into two parts. All parts are drawn to scale on the cover of this publication, and a table at the end lists the thicknesses of the parts. Although the core and mantle are about equal in thickness, the core actually forms only 15 percent of the Earth's volume, whereas the mantle occupies 84 percent. The crust makes up the remaining 1 percent. Our knowledge of the layering and chemical composition of the Earth is steadily being improved by earth scientists doing laboratory experiments on rocks at high pressure and analyzing earthquake records on computers. The Crust Because the crust is accessible to us, its geology has been extensively studied, and therefore much more information is known about its structure and composition than about the structure and composition of the mantle and core. Within the crust, intricate patterns are created when rocks are redistributed and deposited in layers through the geologic processes of eruption and intrusion of lava, erosion, and consolidation of rock particles, and solidification and recrystallization of porous rock. Figure 1. The oceanic crust at the island of Hawaii is about 5 kilometers thick. The thickness of the continental crust under eastern California ranges from 25 kilometers under the Great Valley to 60 kilometers under the Sierra Nevada. By the large-scale process of plate tectonics, about twelve plates, which contain combinations of continents and ocean basins, have moved around on the Earth's surface through much of geologic time. The edges of the plates are marked by concentrations of earthquakes and volcanoes. Collisions of plates can produce mountains like the Himalayas, the tallest range in the world. The plates include the crust and part of the upper mantle, and they move over a hot, yielding upper mantle zone at very slow rates of a few centimeters per year, slower than the rate at which fingernails grow. The crust is much thinner under the oceans than under continents (see figure above). The boundary between the crust and mantle is called the Mohorovicic discontinuity (or Moho); it is named in honor of the man who discovered it, the Croatian scientist Andrija Mohorovicic. No one has ever seen this boundary, but it can be detected by a sharp increase downward in the speed of earthquake waves there. The explanation for the increase at the Moho is presumed to be a change in rock types. Drill holes to penetrate the Moho have been proposed, and a Soviet hole on the Kola Peninsula has been drilled to a depth of 12 kilometers, but drilling expense increases enormously with depth, and Moho penetration is not likely very soon. The Mantle Our knowledge of the upper mantle, including the tectonic plates, is derived from analyses of earthquake waves (see figure for paths); heat flow, magnetic, and gravity studies; and laboratory experiments on rocks and minerals. Between 100 and 200 kilometers below the Earth's surface, the temperature of the rock is near the melting point; molten rock erupted by some volcanoes originates in this region of the mantle. This zone of extremely yielding rock has a slightly lower velocity of earthquake waves and is presumed to be the layer on which the tectonic plates ride. Below this low-velocity zone is a transition zone in the upper mantle; it contains two discontinuities caused by changes from less dense to more dense minerals. The chemical composition and crystal forms of these minerals have been identified by laboratory experiments at high pressure and temperature. The lower mantle, below the transition zone, is made up of relatively simple iron and magnesium silicate minerals, which change gradually with depth to very dense forms. Going from mantle to core, there is a marked decrease (about 30 percent) in earthquake wave velocity and a marked increase (about 30 percent) in density. Figure 2. Cross section of the whole Earth, showing the complexity of paths of earthquake waves. The paths curve because the different rock types found at different depths change the speed at which the waves travel. Solid lines marked P are compressional waves; dashed lines marked S are shear waves. S waves do not travel through the core but may be converted to compressional waves (marked K) on entering the core (PKP, SKS). Waves may be reflected at the surface (PP, PPP, SS). The core was the first internal structural element to be identified. It was discovered in 1906 by R.D. Oldham, from his study of earthquake records, and it helped to explain Newton's calculation of the Earth's density. The outer core is presumed to be liquid because it does not transmit shear (S) waves and because the velocity of compressional (P) waves that pass through it is sharply reduced. The inner core is considered to be solid because of the behavior of P and S waves passing through it. Cross section of the whole Earth, showing the complexity of paths of earthquake waves. The paths curve because the different rock types found at different depths change the speed at which the waves travel. Solid lines marked P are compressional waves; dashed lines marked S are shear waves. S waves do not travel through the core but may be converted to compressional waves (marked K) on entering the core (PKP, SKS). Waves may be reflected at the surface (PP, PPP, SS). Data from earthquake waves, rotations and inertia of the whole Earth, magnetic-field dynamo theory, and laboratory experiments on melting and alloying of iron all contribute to the identification of the composition of the inner and outer core. The core is presumed to be composed principally of iron, with about 10 percent alloy of oxygen or sulfur or nickel, or perhaps some combination of these three elements. This table of depths, densities, and composition is derived mostly from information in a textbook by Don L. Anderson (see Suggested Reading). Scientists are continuing to refine the chemical and mineral composition of the Earth's interior by laboratory experiments, by using pressures 2 million times the pressure of the atmosphere at the surface and temperatures as high as 20000C. The Structure of the Moon The Moon, our fellow-traveler in space, has a diameter half that of the Earth's core, and it revolves around the Earth, as all the planets revolve around the Sun, under the force of gravity. Moonquakes of very low energy are caused by land tides produced by the pull of Earth's gravity, and, from analysis of moonquake data, scientists believe the Moon has two layers: a crust, from the surface to 65 kilometers depth, and an inner, more dense mantle from the crust to the center at 3,700 kilometers. The crust is presumed to be com- posed primarily of rocks containing feldspar, calcium aluminum silicate, and lesser pyrox- ene, iron and magnesium silicate; the crust also contains basalt in the mares, which con- tains less iron and more titanium than earth basalt. The mantle is thought to be made up of calcic peridotite, containing both pyroxene and feldspar. Suggested Reading Anderson, D.L., 1989, Theory of the Earth: Boston, Blackwell Publications, 366 pages. Flint, R.F., and Skinner, B.J., 1977, Physical geology: New York, John Wiley and Sons, 594 pages. Press, Frank, and Siever, Raymond, 1974, Earth: San Francisco, W.H. Freeman, 649 pages. Robertson, E.C., 1966, The interior of the Earth; an elementary description: U.S. Geological Survey Circular 532, 10 pages. Smith, P.J., editor, 1986, The Earth: New York, Macmillan, 248 pages. Yockstick, M.L., 1987, Earthbook -- Encyclopedia of the Earth: Stockholm, Sweden, Esselte Map Service, 327 pages. For sale by the U.S. Government Printing Office Superintendent of Documents, Mail Stop: SSOP, Washington, DC 20402-9328 U.S. Geological Survey Information Services P.O. Box 25286 Denver, CO 80225 As the Nation's principal conservation agency, the Department of the Interior has responsibility for most of our nationally owned public lands and natural and cultural resources. This includes fostering sound use of our land and water resources; protecting our fish, wildlife, and biological diversity; preserving the environmental and cultural values of our national parks and historical places; and providing for the enjoyment of life through outdoor recreation. The Department assesses our energy and mineral resources and works to ensure that their development is in the best interests of all our people by encouraging stewardship and citizen participation in their care. The Department also has a major responsibility for American Indian reservation communities and for people who live in island territories under U.S. administration. Return to General Interest Publications This page is URL: Maintained by Publications Services Last modified 01-14-11 (jmw) Seismic data like the lines pictured on this 3D map help scientists understand the structure of the earth's core. This particular map helps scientists at Chesapeake Energy’s Oklahoma City headquarters choose the best spots to drill. Photograph by Mark ThiessenLearn about the layers inside the Earth, inaccessible to humans.The Earth's interior is composed of four layers, three solid and one liquid—not magma but molten metal, nearly as hot as the surface of the sun. The deepest layer is a solid iron ball, about 1,500 miles (2,400 kilometers) in diameter. Although this inner core is white hot, the pressure is so high the iron cannot melt. The iron isn't pure—scientists believe it contains sulfur and nickel, plus smaller amounts of other elements. Estimates of its temperature vary, but it is probably somewhere between 9,000 and 13,000 degrees Fahrenheit (5,000 and 7,000 degrees Celsius). Above the inner core is the outer core, a shell of liquid iron. This layer is cooler but still very hot, perhaps 7,200 to 9,000 degrees Fahrenheit (4,000 to 5,000 degrees Celsius). It too is composed mostly of iron, plus substantial amounts of sulfur and nickel. It creates the Earth's magnetic field and is about 1,400 miles (2,300 kilometers) thick.River of RockThe next layer is the mantle. Many people think of this as lava, but it's actually rock. The rock is so hot, however, that it flows under pressure, like road tar. This creates very slow-moving currents as hot rock rises from the depths and cooler rock descends.The mantle is about 1,800 miles (2,900 kilometers) thick and appears to be divided into two layers: the upper mantle and the lower mantle. The boundary between the two lies about 465 miles (750 kilometers) beneath the Earth's surface.The crust is the outermost layer of the Earth. It is the familiar landscape on which we live: rocks, soil, and seabed. It ranges from about five miles (eight kilometers) thick beneath the oceans to an average of 25 miles (40 kilometers) thick beneath the continents.Currents within the mantle have broken the crust into blocks, called plates, which slowly move around, colliding to build mountains or rifting apart to form new seafloor.Continents are composed of relatively light blocks that float high on the mantle, like gigantic, slow-moving icebergs. Seafloor is made of a denser rock called basalt, which presses deeper into the mantle, producing basins that can fill with water.Except in the crust, the interior of the Earth cannot be studied by drilling holes to take samples. Instead, scientists map the interior by watching how seismic waves from earthquakes are bent, reflected, sped up, or delayed by the various layers. How can financial brands set themselves apart through visual storytelling? Our experts explain how.Learn MoreThe Motorsport Images Collections captures events from 1895 to today’s most recent coverage.Discover The CollectionCurated, compelling, and worth your time. Explore our latest gallery of Editors’ Picks.Browse Editors' FavoritesHow can financial brands set themselves apart through visual storytelling? Our experts explain how.Learn MoreThe Motorsport Images Collections captures events from 1895 to today’s most recent coverage.Discover The CollectionCurated, compelling, and worth your time. Explore our latest gallery of Editors’ Picks.Browse Editors' FavoritesHow can financial brands set themselves apart through visual storytelling? Our experts explain how.Learn MoreThe Motorsport Images Collections captures events from 1895 to today’s most recent coverage.Discover The CollectionCurated, compelling, and worth your time. Explore our latest gallery of Editors’ Picks.Browse Editors' Favorites Art Lerner-Lam, associate director for seismology, geology, and tectonophysics at the Lamont-Doherty Earth Observatory of Columbia University, explains.Much of our understanding of the basic structure and composition of Earth and the other planets in our solar system is not strenuously debated. We can infer a surprising amount of information from the size, mass and moment of inertia of the planets, all of which can be determined from routine astronomical observations. Measurements of surface chemical composition, either by direct sampling (as has been done on Earth, the moon, and Mars) or through spectroscopic observations, can be used to estimate elemental abundances and the degree of chemical differentiation that occurred as the planets condensed from the solar nebula. Remote observations of the gravitational field can be used to understand how a planet's mass is distributed, whereas the strength and shape of the magnetic field provides some constraint on the structure of a metallic core. The specifics of structure and composition, however, are much more debatable. And it is these details that tell us a much more extensive and ultimately more interesting story about the internal dynamics of the planets and their evolution. As a result, trying to determine them is frontier research in almost all fields of earth and planetary science.Even on Earth, many of these details have to be inferred from remote observations. Because we cannot sample the deep Earth, we must deduce its composition either by looking at the clues hidden in igneous and metamorphic rocks, or by examining proxies for composition and structure such as the three-dimensional variation of the velocity of seismic waves produced by earthquakes and sampled by networks of seismometers on the surface. The late Francis Birch, the eminent Harvard geophysicist, and his colleagues and students worked out the basic methodology that brings these distinct observations together. Birch showed how the stiffness of rocks changes under the extreme conditions of pressure and temperature deep within planets, as well as with chemical composition. Because the speed of seismic waves depends on the stiffness of the medium through which they propagate, it is possible to calculate temperature and composition from maps of seismic velocity. Most current research is based on Birch's work and it has even been extended to the most extreme temperature and pressure conditions of Earth's core. For example, much of our understanding of the large- and small-scale convection patterns driving plate tectonics has come about by using Birch-type proxies for temperature and composition.On supporting science journalismIf you're enjoying this article, consider supporting our award-winning journalism by subscribing. By purchasing a subscription you are helping to ensure the future of impactful stories about the discoveries and ideas shaping our world today.Birch knew, however, that such interpretations should be made cautiously. His seminal paper, published in the Journal of Geophysical Research in 1952, is also famous for its tongue-in-cheek lecture on the uncertainties inherent in extrapolating laboratory and proxy observations to the high pressure and high temperature interiors of planets. Birch provided a small Rosetta stone that enabled future workers to interpret the results his methodology made possible. Thus, when talking about the chemical composition of planetary interiors, ¿certain¿ should be replaced by ¿dubious,¿ ¿positive proof¿ by ¿vague suggestions,¿ and, when talking about Earth's core, ¿pure iron¿ should be replaced by ¿uncertain mixture of all the elements.¿ We obviously know more today than we did 50 years ago, but Birch's words resonate in every classroom and laboratory.How can we improve our understanding of the other planets? Manned and unmanned missions to the moon and Mars deployed seismometers, which provided tantalizing but ultimately limited information before they stopped operating (although the Spirit and Opportunity rovers continue to transmit chemical analyses and pictures of the Red Planet back to Earth). Almost all planetary landing missions now in the design stage include seismological instrumentation and some even include sample return. Hopefully the best science is yet to come. But even on Earth, where thousands of seismometers have been deployed and new experiments, such as the National Science Foundation's ¿EarthScope,¿ are now being conducted, each new observation raises as many questions as it answers. Earth's story has been written, but we're only on the first few chapters.Answer originally published August 23, 2004. Modified from "Physical Geology" by Steven Earle* The previous section described the properties and composition of Earth’s interior, which begs the question: how can we know what conditions are like deep in the Earth? It’s easy to sample the crust through drilling, and mantle material often comes to the surface as magma, but the farthest we have been able to drill into the crust so far is only about 12 km; this for a planet with a radius of 6370 km! So to understand the composition and structure of the Earth’s deep interior, we need to use indirect methods such as seismology. Seismology is the study of vibrations within the Earth. These vibrations are caused by various events, including earthquakes, extraterrestrial impacts, explosions, storm waves hitting the shore, and tidal effects. Of course, seismic techniques have been most widely applied to the detection and study of earthquakes, but there are many other applications, and arguably seismic waves provide the most important information that we have concerning Earth’s interior. Before going any deeper into Earth, however, we need to take a look at the properties of seismic waves. The types of waves that are useful for understanding Earth’s interior are called body waves, meaning that, unlike the surface waves on the ocean, they are transmitted through Earth materials. Imagine hitting a large block of strong rock (e.g., granite) with a heavy sledgehammer. At the point where the hammer strikes it, a small part of the rock will be compressed by a fraction of a millimeter. That compression will transfer to the neighboring part of the rock, and so on through to the far side of the rock, from where it will bounce back to the top — all in a fraction of a second. This is known as a compression wave, and it can be illustrated by holding a loose spring (like a Slinky) that is attached to something (or someone) at the other end. If you give it a sharp push so the coils are compressed, the compression propagates (travels) along the length of the spring and back (Fig. 3.3.1). You can think of a compression wave as a “push” wave — it’s called a P-wave (although the “P” stands for “primary” because P-waves are the first to arrive at seismic stations). In a P-wave the motion of the particles is parallel to the direction of wave propagation. When we hit a rock with a hammer, we also create a different type of body wave, one that is characterized by back-and-forth vibrations (as opposed to compressions). This is known as a shear wave (S-wave, where the “S” stands for “secondary”), and an analogy would be what happens when you flick a length of rope with an up-and-down motion. As shown in Figure 3.3.1, a wave will form in the rope, which will travel to the end of the rope and back. In this case, the motion of the particles is perpendicular to the direction the wave travels. Figure 3.3.1 Representations of a compression wave (P-wave, top) and a shear wave (S-wave, bottom) (Steven Earle, “Physical Geology”). Compression waves and shear waves travel very quickly through geological materials. As shown in Figure 3.2.2, typical P-wave velocities are between 0.5 km/s and 2.5 km/s in unconsolidated sediments, and between 3.0 km/s and 6.5 km/s in solid crustal rocks. Of the common rocks of the crust, velocities are greatest in basalt and granite. S-waves are slower than P-waves, with velocities between 0.1 km/s and 0.8 km/s in soft sediments, and between 1.5 km/s and 3.8 km/s in solid rocks. Figure 3.2.2 Typical velocities of P waves (red) and S waves (blue) in sediments and in solid crustal rocks (Steven Earle, “Physical Geology”). Mantle rock is generally denser and stronger than crustal rock and both P- and S-waves travel faster through the mantle than they do through the crust. Moreover, seismic-wave velocities are related to how tightly compressed a rock is, and the level of compression increases dramatically with depth. Finally, seismic waves are affected by the phase state of rock. They are slowed if there is any degree of melting in the rock. If the material is completely liquid, P-waves are slowed dramatically and S-waves are stopped altogether. Figure 3.3.3 Wave velocities through the different layers of the Earth (left). Enhanced view of wave velocities in the crust and upper mantle (right) (Steven Earle, “Physical Geology”). Accurate seismometers have been used for earthquake studies since the late 1800s, and systematic use of seismic data to understand Earth’s interior started in the early 1900s. The rate of change of seismic waves with depth in the Earth (Fig. 3.3.3) has been determined over the past several decades by analyzing seismic signals from large earthquakes at seismic stations around the world. Small differences in arrival time of signals at different locations have been interpreted to show that: Velocities are greater in mantle rock than in the crust. Velocities generally increase with pressure, and therefore with depth. Velocities slow in the area between 100 km and 250 km depth (called the “low-velocity zone”; equivalent to the asthenosphere). Velocities increase dramatically at 660 km depth (because of a mineralogical transition). Velocities slow in the region just above the core-mantle boundary (the D” layer or “ultra-low-velocity zone”). S-waves do not pass through the outer part of the core. P-wave velocities increase dramatically at the boundary between the liquid outer core and the solid inner core. One of the first discoveries about Earth’s interior made through seismology was in the early 1900s when Croatian seismologist Andrija Mohorovičić (pronounced Moho-ro-vi-chich) realized that at certain distances from an earthquake, two separate sets of seismic waves arrived at a seismic station within a few seconds of each other. He reasoned that the waves that went down into the mantle, traveled through the mantle, and then were bent upward back into the crust, reached the seismic station first because although they had farther to go, they traveled faster through mantle rock (as shown in Figure 3.3.4). The boundary between the crust and the mantle is known as the Mohorovičić discontinuity (or Moho). Its depth is between 60 km and 80 km beneath major mountain ranges, around 30 km to 50 km beneath most of the continental crust, and between 5 km and 10 km beneath the oceanic crust. Figure 3.3.4 Depiction of seismic waves emanating from an earthquake (red star). Some waves travel through the crust to the seismic station (at about 6 km/s), while others go down into the mantle (where they travel at around 8 km/s) and are bent upward toward the surface, reaching the station before the ones that traveled only through the crust (Steven Earle, “Physical Geology”). Our current understanding of the patterns of seismic wave transmission through Earth is summarized in Figure 3.3.5. Because of the gradual increase in density with depth, all waves are refracted toward the lower density, slower velocity material as they travel through homogenous parts of Earth, and thus tend to curve outward toward the surface. Waves are also refracted at boundaries within Earth, such as at the Moho, at the core-mantle boundary (CMB), and at the outer-core/inner-core boundary. S-waves do not travel through liquids — they are stopped at the CMB — and there is an S-wave shadow on the side of Earth opposite a seismic source. The angular distance from the seismic source to the shadow zone is 103° on either side, so the total angular distance of the shadow zone is 154°. We can use this information to infer the depth to the CMB. P-waves do travel through liquids, so they can make it through the liquid part of the core. Because of the refraction that takes place at the CMB, waves that travel through the core are bent away from the surface, and this creates a P-wave shadow zone on either side, from 103° to 150°. This information can be used to discover the differences between the inner and outer parts of the core. Figure 3.3.5 Patterns of seismic waves moving through Earth’s interior. Since S waves do not pass through the liquid outer core, a shadow zone is created on the opposite side from the original disturbance (Steven Earle, “Physical Geology”). Using data from many seismometers and hundreds of earthquakes, it is possible to create a two- or three-dimensional image of the seismic properties of part of the mantle. This technique is known as seismic tomography, and an example of the result is shown in Figure 3.3.6. Figure 3.3.6 Seismic tomography image showing the Pacific Plate (blue) subducting beneath Tonga (Steven Earle, “Physical Geology”). The Pacific Plate subducts beneath Tonga and appears in Figure 3.3.6 as a 100 km thick slab of cold (blue-colored) oceanic crust that has pushed down into the surrounding hot mantle. The cold rock is more rigid than the surrounding hot mantle rock, so it is characterized by slightly faster seismic velocities. There is volcanism in the Lau spreading center and also in the Fiji area, and the warm rock in these areas has slower seismic velocities (yellow and red colors). Seismic waves provide us with the structure of the inner Earth, but what about it’s other properties? In terms of composition, there are several lines of evidence pointing to a core composed mostly of iron and nickel. Wave properties suggest the core is composed of an element with an atomic number around 25 (iron has an atomic number of 26). Aside from iron, all of the other elements with an atomic number close to 25 are too rare to make up the core. If the Earth was formed through the accretion of smaller bodies such as meteorites, we would expect the composition of Earth to be similar to the composition of meteorites. Meteorites are mostly iron and nickel, but in higher proportions than the Earth’s crust. This suggests that most of this heavy iron and nickel from the meteorites must have sunk to the Earth’s center as the planet was forming. However, the core is not dense enough to be pure iron and nickel; it it about 10% below the predicted density if that was the case. This is why scientists believe the core is composed of about 10% sulfur, oxygen, and hydrogen. Finally, if the Earth’s magnetic field comes from the fluid outer core, the outer core must contain iron. In terms of the temperatures, we can calculate the melting points of these materials over the range of pressures that they would experience in the inner Earth, and then infer the temperatures that would allow these elements to exist in their solid or liquid forms. *”Physical Geology” by Steven Earle used under a CC-BY 4.0 international license. Download this book for free at definitionthe uppermost layer of the Earth, ranging in thickness from about 5 km (in the oceans) to over 50 km (on the continents) (3.2) the middle layer of the Earth, dominated by iron and magnesium rich silicate minerals and extending for about 2900 km from the base of the crust to the top of the core (3.2) the study of vibrations within the Earth (3.3) a seismic wave that travels through rock (e.g., a P-wave or an S-wave) (3.3) an igneous (formed from cooling magma) rock that comprises much of the continental crust (3.2) a seismic body wave that is characterized by deformation of the rock in the same direction that the wave is propagating (compressional vibration) (3.3) a seismic body wave that is characterized by deformation of the rock perpendicular to the direction that the wave is propagating (3.3) a volcanic rock that makes up much of the oceanic crust (3.2) unconsolidated particles of mineral or rock that settle to the seafloor (12.1) the change of state between a solid, liquid, or gas (8.1) the part of the mantle, from about 100 to 200 km below surface, within which the mantle material is close to its melting point, and therefore relatively weak (3.2) the metallic interior part of the Earth, extending from a depth of 2900 km to the center (3.2) the layer of the inner Earth extending 2300 km from the top of the inner core to the bottom of the mantle, composed of fluid metal alloys (3.2) the solid metal mass at the center of the Earth, extending 1200 km from the center (3.2) the boundary between the crust and the mantle (3.3) mass per unit volume of a substance (e.g., g/cubic cm) (6.3) when part of a plate is forced beneath another plate along a subduction zone (4.3) the Earth’s crust underlying the oceans (as opposed to continental crust) (3.2) the process by which solid celestial bodies are added to existing bodies during collisions (3.1) a fragment of either stony or metallic debris in space (12.5) The Earth’s interior comprises different layers with varying physical properties and compositions. The layered structure of the Earth has been interpreted based on the studies of seismic waves, magnetic fields, meteorites, and other physical properties. Base on the chemical composition, the internal structure of the earth is divided majorly into crust, mantle and core. The core is further sub-divided into outer core and inner core. The crust is the outermost layer of the Earth above the Mohorovicic discontinuity. It is defined based on composition. Thickness of the crust ranges from about 3 km at some oceanic ridges to about 70 km in collisional orogens. It consists of the sial or both the sial and the sima. The crust represents less than 0.1% of the Earth’s total volume. Sial is a term used in petrology to refer to the upper layer of the Earth’s crust. This term is based on silica and alumina, the principal components of the rocks in this layer. Sial is known for being the primary source of granitic magma and distinguishes the upper continental crust. Sima (an acronym for silica + magnesia) is a petrologic term for the lower layer of the Earth’s crust. It consists of rocks that are abundant in silica and magnesia. Sima is compositionally similar to the oceanic crust and the lower section of the continental crust beneath the sial. Further, scientists divide the Crust into two types: continental crust (beneath land masses) and oceanic crust (beneath oceans). Continental crust is a type of the Earth’s crust which underlies the continents and the continental shelves. It ranges in thickness from about 25 km to as much as 70 km under mountain ranges, averaging ~40 km. The density of the continental crust averages ~2.7 g/cm3. It has the planet’s oldest rocks (billions of years old). The velocities of compressional seismic waves through its average ~6.5 km/s and are less than ~7.0 km/sec. The oceanic crust is a type of Earth’s crust that lies beneath the ocean basins. It is generally 5-10 km thick and denser than the continental crust, with a density of 2.9 g/cm3. Compressional seismic-wave velocities travels through it at 4-7.5 km/sec. Oceanic crusts are geologically young, with an age of 200 million years or less. The mantle is a layer beneath the Earth’s crust and above the core. It extends from about 70-km to 2,900-km in depth. It contains about 80% of the Earth’s volume. The boundary between the crust and mantle shows a change in chemical composition. The mantle comprises more magnesium and iron than the minerals found in either the continental or oceanic crust. It can be divided into divided into the upper and lower mantle. Upper mantle is part of the mantle which lies between the Moho to 660 km depth. It includes the lower part of the Iithosphere and the upper part of the asthenosphere. The density of upper mantle ranges from 3.4 g/cm3 to 4.4 g/cm3 with increasing depth. Likewise, P-wave velocity increases from about 8 to 11 km/sec with depth and S-wave velocity increases from about 4.5 to 6 km/sec with depth. Composition of upper mantle is presumed to be peridotitic in composition. The region from the 410-km to the 660-km discontinuity within upper mantle is known as the transition zone. These two discontinuities are result of two important solid-state transformations: from olivine to wadsleyite at 410 km and from spinel to perovskite + magnesiowustite at 660 km. Lower mantle is lower part of the mantle that lies below a depth of about 660 km to 2900 km at the core-mantle boundary. It is characterized by rather constant increases in velocity and density in response to increasing hydrostatic compression. The density of upper mantle ranges from ~4.4 g/cm3 to ~5.6 g/cm3 with increasing depth. The velocity of compressional seismic waves increases from ~10.7 km/s to ~13.7 km/s. Core is the central zone or nucleus of the Earth’s interior, below the Gutenberg discontinuity at a depth of 2900 km. The core comprises about 16% of the Earth’s volume and 32% by mass. It is divided into an inner core and an outer core, with a transition zone between. Since only compressional waves propagate in the outer core, it is a fluid. The inner core, having a radius of approximately 1300 km, is solid, as shear waves have been observed to propagate through it. The magnetic field originates within the core. The Outer Core is upper zone of the Earth’s core. It is extending from a depth of 2900 km to 5200 km. It is presumed to be liquid because it sharply reduces compressional-wave velocities and does not transmit shear waves. The density of outer core ranges from 9.9 to 12.2 g/cm3. The outer core is the source of the principal geomagnetic field. The Inner Core is the central part of the Earth’s core. It is extending from a depth of about 5200 km to the center (6371 km) of the Earth. Its radius is about one third of the whole core. The inner core is considered solid, as evidenced by the observation of S waves that are propagated in it, and because compressional waves travel noticeably faster through it than through the outer core. The density of inner core ranges from 12.8 to 13.1 g/cm3. The physical (or mechanical) properties of a material tell us how it responds to force, how weak or strong it is, and whether it is a liquid or a solid. Based on physical properties internal structure of the Earth is divided into – (1) Lithosphere, (2) Asthenosphere, (3) Mesosphere, (4) Outer Core and (5) Inner Core. Lithosphere (“rock sphere”) includes the crust and part of the upper mantle. It is solid, strong, and rigid outer layer of a planet. Earth’s lithosphere varies greatly in thickness, from as little as 10 km in some oceanic areas to as much as 300 km in some continental areas. Asthenosphere (“weak sphere”) is the layer of the Earth below the lithosphere, which is weak and in which isostatic adjustments take place. It is extending from the base of the Iithosphere to the 660-km discontinuity, is by comparison a weak layer that readily deforms by creep. It is a part of the upper mantle where temperature and pressure are just right so that part of the material melts, or nearly melts. Under these conditions, rocks lose much of their strength and become soft and plastic and flow slowly. It is presumed that magmas may be generated in this zone. Here seismic waves are strongly attenuated. The boundary between the asthenosphere and the overlying lithosphere is mechanically distinct but does not correspond to a fundamental change in chemical composition. The boundary is simply a major change in the rock’s mechanical properties. Mesosphere (“middle sphere”) is the region between the asthenosphere and the core. The rock below the asthenosphere is stronger and more rigid than in the asthenosphere. It is so because the high pressure at this depth offsets the effect of high temperature, forcing the rock to be stronger than the overlying asthenosphere. Mohorovicic discontinuity is defined as the boundary or sharp seismic-velocity discontinuity that separates the Earth’s crust from the subjacent mantle. It marks the level in the Earth at which P-wave velocities change abruptly from 6.7-7.2 km/sec (in the lower crust) to 7.6-8.6 km/sec or average 8.1 km/sec (at the top of the upper mantle). The depth of Mohorovicic discontinuity ranges from about 5-10 km beneath the ocean floor to about 40 km below the continents. In some cases, it may reach up to 70 km under mountain ranges. Mohorovicic discontinuity probably represents a chemical change from basaltic or simatic materials above to peridotitic or dunitic materials below, rather than a phase change (basalt to eclogite). However, the discontinuity should be defined by seismic velocities alone. It is variously estimated to be between 0.2 and 3 km thick. It is named in honour of its discoverer, Andrija Mohorovicic (1857-1936). Gutenberg discontinuity is the seismic-velocity discontinuity at 2900 km depth. It marks the mantle-core boundary, at which the velocities of P waves are reduced and S waves disappear. It probably reflects the change from a solid to a liquid phase as well as a change in composition. Gutenberg discontinuity is named after Beno Gutenberg, a seismologist. The low-velocity zone (LVZ) in the upper mantle is characterized by low seismic-wave velocities, high seismic energy attenuation, and high electrical conductivity. The bottom of the LVZ, sometimes referred to as the Lehmann discontinuity, has been identified from the study of surface wave and S-wave data in some continental areas (Figure 4.1) (Gaherty and Jordan, 1995). This discontinuity, which occurs at depths of 180-220 km, appears to be thermally controlled and at least in part reflects a change from an anisotropic lithosphere to a more isotropic asthenosphere. Because of the dramatic drop in S-wave velocity and increase in attenuation of seismic energy, it would appear that partial melting must contribute to producing the LVZ. The zone from the 4l0-km to the 660-km discontinuity is known as the transition zone. The transition zone is the part of the upper mantle where two major seismic discontinuities – one at 410 km and the other at 660 km occur. High-pressure experimental results indicate that at about 14 Gpa equivalent to 410 km burial depth in the Earth, the Mg-rich olivine breaks to a high-pressure phase known as wadsleyite (beta phase). No chemical composition change occurs during this phase change or other phase changes. Mantle olivine (F090) completely transforms to wadsleyite over a < 300 Mpa pressure range at appropriate temperatures for the 410-km discontinuity (~ 1000 °C) (Ita andStixrude, 1992). Similar to the 410-km discontinuity, it appears that a phase change in Mg2Si04 is responsible for the 660-km discontinuity (Christensen, 1995). High-pressure experimental studies suggest that spinel transforms to a mixture of perovskite and magnesiowustite at a pressure of about 23 Gpa, and can explain both the changes in seismic velocity and density at this boundary if the rock contains 50-60 per cent spinel. Magnesiowustite and Mg-perovskite are high-density and appear to comprise most of the lower mantle. Did you know that the lower mantle is mostly made up of two extremely dense minerals, Mg perovskite and magnesiowustite? It’s fascinating how these minerals can withstand such intense pressure and heat deep within the Earth! The D” layer is a region of the mantle exists within a few hundred kilometres of the core where seismic velocity gradients are anomalously low (Young and Lay, 1987; Loper and Lay, 1995). A small temperature gradient (1-3 °C/km) can conduct heat from the core into the D” layer. Seismic wave diffraction by the core causes poor resolution in this layer, leading to limited knowledge of its structure details. Estimates of the thickness of the D” layer suggest that it ranges from 100 to about 500 km. The final frontier isn’t space: It’s the Earth itself. We’ve sent people to the moon, robots to Mars and the New Horizons space probe 3.26 billion miles from Earth to snap photos of Pluto, while just 4,000 miles beneath our feet, unfathomable heat and pressure keep the center of the Earth tantalizingly out of reach. But scientists have been able to puzzle out what’s inside the Earth — including olive-green crystals and a roiling sea of melted iron — by studying meteorites, volcanic eruptions and the seismic waves from earthquakes.“We go and explore other planets, but in many ways, going inside the Earth and figuring out what's inside the Earth is actually technologically harder than going into space,” says Vedran Lekić, a seismologist at the University of Maryland.There are four main layers to the Earth: crust, mantle, outer core and inner core, along with transition zones between these layers. The world we know lies on tectonic plates making up the Earth’s crust, which varies in thickness from three miles to over 40. Beneath the crust lies the mantle, the layer of rock making up 84 percent of the Earth’s volume. The rocks in the upper mantle are white-hot, but if you could cool them to room temperature, they’d be a speckly olive green thanks to the mineral olivine — you might know it as the August birthstone peridot. “I think the mid-upper mantle would be gorgeous, because it would be olivine green, like 60 percent, and it would also have garnets, these beautiful red cubic minerals,” says Wendy Mao, a mineral physicist at Stanford University.Deeper in the mantle, heat and pressure reconfigure the atoms making up olivine into two new minerals, bridgmanite and ferropericlase, which are brownish-orange and yellow at room temperature. Beneath the rocky mantle, there’s an outer core of churning liquid iron (and a little nickel) surrounding an inner core of solid iron (again with some nickel) that’s about 70 percent the size of the moon. The center of the Earth is almost as hot as the surface of the Sun, about 9,800 degrees Fahrenheit, with pressure that makes the compressive forces at the ocean floor look like child’s play.But we haven’t been to any of these places inside the Earth. We haven’t seen them. We haven’t sent cameras or probes to bring back samples of iron glowing like coal embers. So, how do we know what’s there?Earthquakes Reveal StructureWhen the tectonic plates that make up the Earth’s crust move past each other, they sometimes catch and break. This breakage, along with the waves of energy that come with it, is called an earthquake. The seismic waves unleashed, sound waves too low for us to hear, travel far from the source and can tell scientists when and where an earthquake has occurred.Detecting seismic waves isn’t new. The Chinese scientist Zhang Heng built an early seismometer nearly 2,000 years ago. But in 1889, scientists had a breakthrough in using them to understand our planet. That’s when a German researcher near Berlin detected an earthquake, but there was a problem: There weren’t any earthquakes nearby that day. It turns out, there had been an earthquake — in Japan, and its seismic waves reached Germany more than an hour after it struck. It marked a pivotal point in modern seismology.Lekić likens seismology to the way we use X-rays to see inside the human body — the different densities of our muscles, organs, and bone mean that the X-rays travel through them (or get deflected by them) in different ways. “We can’t send X-rays through the Earth, because the X-rays won’t make it all the way through,” he says. “Instead, we use seismic waves.”Read more: 20 Things You Didn't Know About Inner EarthEarthquakes send ground vibrations throughout our planet, and they travel across the world in different ways. Seismometers record these vibrations, which can offer a glimpse of what’s below, explains Lekić: “Then we can use computer modeling to try to make images of what’s inside the Earth, and those images tell us essentially how quickly seismic waves travel through different parts of the Earth.”Seismologists interpret data from earthquakes and even simulate seismic activity with air guns and explosions, and their work has shown that the Earth’s interior has different layers, some of which are easier for seismic waves to travel through than others. They can even tell us the densities of these layers. Then, a different branch of scientists called mineral physicists take it from there to determine what actually makes up those layers.Reading the Minerals“There are astronomical and cosmic chemical constraints on understanding what are the likely building blocks of our planet,” says Andrew Campbell, a mineral physicist at the University of Chicago. Basically, the distribution of elements in the Earth should roughly match the elements present in meteorites and the Sun.“[The material] we have available in the solar system that we could make a planet from, it should be the same kind of stuff that's in the Sun,” Campbell explains. This is because chemical reactions inside stars produce the elements that make up planets like the Earth. By looking at the wavelengths of light that the sun gives off and comparing those wavelengths to light bouncing off known elements, scientists can glean the chemical makeup of the sun. The light shows there's a lot of silicon, oxygen, magnesium and iron, along with other elements like potassium and calcium, Campbell says. The relative amounts of elements in the sun is similar to what we see in certain primitive meteorites, he adds, which "reinforces our understanding that these primitive meteorites represent the building blocks from which we can assemble terrestrial planets. And that includes Earth."These clues from beyond the Earth can give us a sense of the key players in the planet's composition, as can volcanic rocks that contain pieces of the Earth's mantle. Then, mineral physicists like Campbell and Mao figure out how these elements must be distributed in order for the seismology data to add up."I would argue this is how we understand that the Earth's core is rich in iron and not some other heavy element," Campbell says. Seismology tells us the Earth has a solid, dense core. Since the sun and meteorites contain more iron than other heavy elements like cobalt, nickel, or chromium, he says, "we know that iron is a really big part of the building blocks from which we're assembled." Mineral physicists also find ways to mimic conditions within the Earth that affect the materials that make it up. For instance, Mao has done studies to learn how the iron at the Earth's core could have settled there long ago, slowly working its way through the rocky mantle. To simulate the intense pressure within the Earth, she uses what's called a diamond anvil cell. "We take two diamonds and then basically cut off the tip, take that small surface area, put a little sample in between them, and then squeeze the backs," she says. Since the force is concentrated on such a small surface area, about the width of a human hair, it's compounded in a way comparable to the pressure deep within the planet."One of the most exciting, but also frustrating, parts about these studies is that we can't go down to that depth," says Mao. "There's theory that makes sense, that's consistent with experimental data, because we can try to mimic what we think the conditions are, albeit micron-sized."And since scientists haven't found a way to travel to the center of the Earth, the collaboration between these different disciplines is critical to understanding what lies beneath our feet. "Without kind of working together," says Lekić, "you can't really understand the planet."