Geophysics

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Contents

Unit 1 – Introduction to Geophysics

1.1 Overview of geophysics and its applications

Geophysics is the study of Earth using quantitative physical methods. It combines principles from physics, geology, and mathematics to investigate everything from earthquakes to magnetic fields, helping us understand our planet's interior structure, surface processes, and available resources. Because geophysics spans so many domains, it plays a central role in resource exploration, hazard assessment, environmental monitoring, and climate science.

Geophysics: Definition and Scope

Definition and Areas of Study

Geophysics applies the laws of physics and mathematical tools to study the Earth. It draws on geology for context about Earth materials and structures, and on physics for the governing equations that describe how energy and matter behave. The major subdisciplines are:

- Seismology: Study of elastic waves that propagate through the Earth, generated by earthquakes or artificial sources.
- Geodesy: Study of the Earth's shape, gravity field, and surface deformation, often using satellite-based techniques.
- Geodynamics: Study of the forces and processes that drive plate tectonics and mantle convection.
- Geomagnetism: Study of the Earth's magnetic field, its origin in the core, and how it changes over time.
- Geothermics: Study of the Earth's internal heat flow and temperature distribution.
- Geophysical fluid dynamics: Study of fluid motion in the Earth's core, mantle, oceans, and atmosphere.

Scope and Scale of Geophysical Investigations

Geophysical studies cover an enormous range of spatial and temporal scales. Spatially, investigations extend from the inner core (about 5,150 km deep) all the way out through the atmosphere. Temporally, they span from milliseconds (the travel time of a seismic wave between two sensors) to billions of years (the thermal and compositional evolution of the planet).

Geophysical methods apply across multiple Earth domains: the solid Earth, oceans, atmosphere, and cryosphere (glaciers and ice sheets). This breadth is what makes geophysics valuable for understanding Earth as a complex, interconnected system rather than a collection of isolated parts.

Applications of Geophysics

Resource Exploration and Characterization

Geophysical methods help locate and characterize subsurface resources, including:

- Oil and gas reservoirs: Seismic reflection surveys image sedimentary layers and structural traps where hydrocarbons accumulate.

- Mineral deposits: Gravity, magnetic, and electromagnetic surveys detect density, magnetization, and conductivity contrasts associated with ore bodies (e.g., gold, copper, iron ore).
- Groundwater resources: Electrical resistivity and seismic refraction methods map aquifer geometry and water table depth.
- Geothermal energy sources: Heat flow measurements and resistivity surveys identify areas of elevated subsurface temperatures suitable for energy extraction.

Geophysical data guide drilling and extraction strategies, which helps optimize resource recovery while minimizing environmental impact.

Environmental Monitoring and Assessment

Environmental geophysics focuses on detecting and tracking subsurface contamination and pollution. Common problems include:

- Groundwater contamination by industrial pollutants or agricultural chemicals
- Soil pollution by heavy metals or organic contaminants
- Subsurface waste in landfills or underground storage facilities

Techniques like electrical resistivity tomography (ERT) and ground-penetrating radar (GPR) can map contaminant plumes in the subsurface without drilling. These methods also monitor the progress of remediation efforts over time, helping evaluate whether cleanup strategies are working and guiding sustainable land-use decisions.

Natural Hazard Assessment and Mitigation

Geophysical techniques are central to understanding and preparing for natural hazards:

- Earthquakes: Seismic monitoring networks detect and locate earthquakes in real time, while detailed studies characterize fault zones and estimate future seismic potential.
- Volcanic eruptions: Monitoring of seismic activity, ground deformation (via GPS and InSAR), and gas emissions helps forecast eruptions.
- Landslides: Geophysical surveys assess slope stability and track ground movement before failure occurs.
- Tsunamis: Numerical models simulate wave propagation and coastal inundation to define risk zones.

These data feed into hazard zonation maps, early warning systems, and building codes. The goal is to reduce the impact of natural disasters through better planning and faster response.

Climate Change and Earth System Science

Geophysics contributes to climate science by providing observations and models of key processes:

- Earth's energy balance: Measurements of radiative transfer and greenhouse gas concentrations constrain how much energy the planet retains.
- Ocean circulation: Geophysical observations track heat transport and CO_2 uptake by the oceans.
- Ice sheet dynamics: Satellite gravimetry (e.g., the GRACE mission) measures ice mass loss and its contribution to sea-level rise.

- Permafrost thaw: Monitoring tracks the release of greenhouse gases and associated landscape changes in polar regions.

Satellite measurements of Earth's gravity field and sea surface height provide continuous, global-scale monitoring of climate-related changes. Geophysical models then simulate future scenarios to inform mitigation and adaptation strategies.

Archaeological and Cultural Heritage Investigations

Geophysical methods allow archaeologists to detect and map buried structures without excavation, preserving fragile sites. Commonly used techniques include GPR, magnetic gradiometry, and ERT. These can identify:

- Foundations of ancient buildings and settlements
- Burial sites and cemeteries
- Middens (refuse deposits) that reveal information about daily life
- Prehistoric landscapes and land-use patterns

Geophysical surveys guide targeted excavations so that digging happens only where it's most informative, minimizing damage to cultural heritage. Combining geophysical data with archaeological findings gives a more complete picture of past societies and how they interacted with their environment.

Interdisciplinary Nature of Geophysics

Integration of Principles and Methods from Various Disciplines

Geophysics sits at the intersection of several fields:

- Physics provides the fundamental laws governing wave propagation, gravity, electromagnetism, and heat transfer.
- Geology supplies knowledge of Earth materials, rock structures, and geological history.
- Mathematics underpins quantitative analysis, inverse theory, and numerical modeling.
- Chemistry contributes understanding of material properties and reactions relevant to the Earth's interior.
- Computer science enables data processing, visualization, and large-scale numerical simulations.

This interdisciplinary foundation is what allows geophysicists to tackle problems that no single discipline could solve alone.

Collaboration with Other Earth Science Disciplines

Geophysicists work closely with geologists. Physical methods like seismic imaging reveal subsurface structure that complements surface observations from rock outcrops and drill cores. Joint interpretation of both data types produces a more complete picture of what lies beneath the surface.

Geophysical data also combine with geochemical and petrological analyses. For example, seismic velocities constrain the density and rigidity of deep Earth materials, while geochemistry constrains their composition. Together, these approaches improve our understanding of how rocks form and deform at depth.

Geophysics connects to atmospheric and ocean sciences as well. The solid Earth, oceans, and atmosphere interact continuously. Coupled models that simulate these interactions are essential for studying processes like climate change, where ice sheet behavior, ocean circulation, and atmospheric composition all influence each other.

Reliance on Advancements in Instrumentation and Computational Methods

Progress in geophysics depends heavily on technological advances from engineering and computer science. Better sensors, such as broadband seismometers and superconducting gravimeters, improve the quality and resolution of measurements. High-performance computing makes it possible to process massive datasets and run simulations of complex Earth processes that would have been impossible a few decades ago.

More recently, data science and machine learning techniques have opened new approaches to analyzing and interpreting geophysical data, from automated earthquake detection to noise suppression in seismic records. Collaboration with engineers also drives innovation in data acquisition platforms, such as autonomous underwater vehicles used for seafloor mapping.

Geophysical Methods for Earth Investigation

Seismology

Seismology uses the propagation of elastic waves to image the Earth's interior and study dynamic processes. These waves are generated by earthquakes or artificial sources (e.g., explosions, vibroseis trucks) and recorded by seismometers at the surface.

Two main types of seismic body waves travel through the Earth's interior:

- P-waves (primary/compressional): Travel fastest and arrive first at a seismometer. They can pass through solids and liquids.
- S-waves (secondary/shear): Slower than P-waves and cannot travel through liquids, which is how we know the outer core is liquid.

Seismic tomography uses travel-time data from many earthquakes and stations to construct 3D images of the Earth's interior velocity structure. These images reveal features like the core-mantle boundary, subducting slabs, and mantle plumes.

Seismology is also used to study earthquake source mechanisms, plate boundary dynamics, mantle convection patterns, and core structure and rotation.

Geodesy

Geodesy measures the Earth's shape, gravity field, and surface deformation, primarily using satellite-based techniques.

- GPS (Global Positioning System) measures precise positions and velocities of points on the Earth's surface. GPS data track plate motions (typically a few cm/year), crustal deformation near faults, and post-glacial rebound.
- InSAR (Interferometric Synthetic Aperture Radar) compares satellite radar images taken at different times to map surface deformation with millimeter-scale precision. It's applied to earthquake deformation, volcanic inflation, and land subsidence caused by groundwater extraction.

- Satellite gravimetry missions like GRACE (Gravity Recovery and Climate Experiment) measure temporal variations in the Earth's gravity field. These data monitor ice sheet mass balance, groundwater depletion, and large-scale solid Earth deformation.

Geomagnetism

The Earth's magnetic field originates from convective motions of electrically conductive iron-rich fluid in the outer core. This process is called the geodynamo.

Paleomagnetic studies analyze the magnetization recorded in rocks and sediments, which preserves a record of past geomagnetic field orientations and intensities. This record is used to reconstruct past plate tectonic motions and to study geomagnetic reversals, events where the north and south magnetic poles swap.

Modern magnetic surveys and satellite missions (e.g., the ESA Swarm constellation) map the spatial distribution and temporal variations of the geomagnetic field. Magnetic anomalies at the surface help explore crustal structure and locate mineral resources with magnetic signatures.

Geothermics

Geothermics studies how heat moves through and out of the Earth. Borehole temperature measurements combined with thermal conductivity data allow calculation of heat flow (measured in $\mathrm{mW/m^2}$). These heat flow values constrain the thermal structure of the lithosphere and deeper mantle.

Surface expressions of geothermal activity, such as hot springs and geysers, reveal where fluid circulation transfers heat efficiently through the crust. These systems can be harnessed for electricity generation and direct heating.

Numerical models of mantle convection incorporate geothermal data to simulate the thermal evolution of the Earth over geological time.

Geophysical Fluid Dynamics

Geophysical fluid dynamics applies the physics of fluid motion to Earth systems at scales ranging from laboratory experiments to the entire planet. Fluid dynamical processes are central to:

- Mantle convection and plate tectonics: Slow, viscous flow in the mantle drives plate motions at the surface.
- Core dynamics: Turbulent convection in the liquid outer core generates the geomagnetic field.
- Ocean circulation: Large-scale currents transport heat and influence climate.
- Atmospheric circulation: Wind patterns, jet streams, and weather systems are all governed by fluid dynamics on a rotating sphere.

Numerical simulations and laboratory experiments investigate these phenomena across scales, from small convection cells to planetary-scale circulation patterns. Geophysical fluid dynamics provides the framework for understanding how the solid Earth, oceans, and atmosphere couple together as parts of one system.

1.2 Earth's structure and composition

Earth's structure and composition form the foundation of geophysics. Understanding what the planet is made of and how it's organized at depth is essential for interpreting seismic data, explaining plate tectonics, and modeling processes like volcanism and magnetic field generation.

Seismic waves are the primary tool for probing Earth's interior. Because P-waves and S-waves respond differently to solid and liquid materials, they act as remote sensors that map out boundaries between layers thousands of kilometers below the surface.

Earth's Interior Layers

Major Layers and Their Depths

The Earth is divided into four main layers, each with distinct physical and chemical properties.

- Crust: The outermost layer, ranging from about 5 km thick beneath the oceans to roughly 70 km thick under continental mountain belts. It's composed of solid rocks and minerals.
- Mantle: Extends from the base of the crust down to about 2,900 km depth. The mantle is made of hot, dense silicate rock. In the upper mantle (roughly 100–200 km depth), a partially molten zone called the asthenosphere allows tectonic plates to move.
- Outer core: A liquid layer from approximately 2,900 to 5,100 km depth, composed primarily of iron-nickel alloy.
- Inner core: The innermost layer, from about 5,100 km to the Earth's center at 6,371 km. It's solid iron-nickel alloy under immense pressure and temperature.

Temperature and Pressure Changes with Depth

Both temperature and pressure increase with depth. This gradient drives the different physical states of each layer.

- In the outer core, temperatures are extremely high (roughly 4,000–5,000°C), but the material remains liquid because the pressure, while enormous, isn't sufficient to force it into a solid state at those temperatures.
- In the inner core, the pressure is so great (estimated at over 360 GPa) that it forces the iron-nickel alloy into a solid state despite temperatures comparable to the surface of the Sun (~5,000–6,000°C).

The key takeaway: it's the balance between temperature and pressure that determines whether a given layer is solid, liquid, or able to flow plastically.

Composition and Properties of Earth's Layers

Density and Composition Variations

Density increases systematically from the surface to the center, reflecting changes in both composition and pressure.

Layer	Density (g/cm³)	Primary Composition
Crust	2.7–3.0	Silicate rocks (O, Si, Al, K)
Mantle	3.3–5.7	Silicate rocks (Mg, Fe, Ca, Al)
Outer core	9.9–12.2	Liquid Fe-Ni alloy
Inner core	12.8–13.1	Solid Fe-Ni alloy

The crust is dominated by lighter elements like silicon and aluminum. Moving deeper, iron and magnesium become increasingly dominant. By the core, the composition is overwhelmingly iron and nickel, with minor amounts of lighter elements (sulfur, oxygen, or silicon, depending on the model).

Physical State and Behavior of Materials

- The crust and uppermost mantle (together called the lithosphere) behave as rigid, brittle solids. This is the layer that breaks during earthquakes.
- The asthenosphere (upper mantle, roughly 100–200 km depth) is solid but can deform plastically over geological timescales. This ductile flow is what allows tectonic plates to move.
- The lower mantle is also solid but convects very slowly, driving large-scale mantle circulation.
- The outer core is liquid. Convection currents in this electrically conducting fluid generate Earth's magnetic field through a process called the geodynamo.
- The inner core is solid. It may rotate slightly faster than the rest of the Earth, though this is still debated.

These physical states directly control processes like plate tectonics, volcanism, and seismic wave propagation.

Seismic Waves and Earth's Structure

Types of Seismic Waves and Their Propagation

Seismic waves, generated by earthquakes or artificial sources (like explosions used in exploration), travel through the Earth and carry information about what they've passed through.

- P-waves (primary waves) are compressional: they push and pull material in the direction of travel. They propagate through both solids and liquids.
- S-waves (secondary waves) are shear waves: they move material perpendicular to the direction of travel. They can only propagate through solids because liquids don't support shear stress.

The velocity of both wave types depends on the density and elastic properties of the material. When waves cross a boundary between layers of different properties, they refract (bend) or reflect, just like light passing between air and water.

Evidence for Earth's Internal Structure

Several key observations from seismology constrain the depth and nature of each layer:

1. The Mohorovičić discontinuity (Moho): A sharp increase in seismic velocity marks the boundary between the crust and mantle. Andrija Mohorovičić identified it in 1909 from the arrival times of seismic waves at different distances from an earthquake.
2. S-wave shadow zone: S-waves are not detected at angular distances greater than about 104° from an earthquake source. This tells us the outer core is liquid, since S-waves can't travel through it.
3. P-wave shadow zone: P-waves are absent between roughly 104° and 140° from the source. This results from refraction of P-waves as they enter the liquid outer core, bending them away from that zone. Faint P-wave arrivals within the shadow zone were later explained by reflections off the solid inner core.
4. Seismic tomography: By analyzing travel times from thousands of earthquakes recorded at stations worldwide, geophysicists build 3D velocity models of the interior. These images reveal heterogeneity within layers, such as hot upwellings (mantle plumes) and cold downwellings (subducting slabs).

Rocks and Minerals of the Crust and Mantle

Rock Types and Formation Processes

The crust and upper mantle are composed of three fundamental rock types, classified by how they form:

- Igneous rocks form from the cooling and solidification of magma or lava. Basalt (fine-grained, from rapid cooling at the surface) dominates oceanic crust, while granite (coarse-grained, from slow cooling at depth) is characteristic of continental crust.
- Sedimentary rocks form from the accumulation and lithification of sediments produced by weathering and erosion. Sandstone (from sand-sized grains) and limestone (often from biological carbonate material) are common examples.
- Metamorphic rocks form when pre-existing rocks are transformed by elevated temperature and pressure without fully melting. Gneiss (from granite or sedimentary protoliths) and marble (from limestone) are typical examples.

Common Rock-Forming Minerals

- Silicate minerals are by far the most abundant in the crust and upper mantle. Key examples include quartz, feldspar, mica, pyroxene, and olivine. Olivine and pyroxene dominate the upper mantle, while quartz and feldspar are more common in crustal rocks.
- Carbonate minerals like calcite and dolomite are the main constituents of sedimentary limestones and dolostones. They often form through biological activity (shells, coral) or chemical precipitation from seawater.
- Oxide minerals such as magnetite and hematite are important in some igneous and metamorphic rocks. Magnetite is particularly relevant in geophysics because it carries a record of Earth's magnetic field at the time the rock formed.

Factors Influencing Rock and Mineral Distribution

The types of rocks and minerals you find in a given region depend on its geological context:

- Tectonic setting is a major control. Mid-ocean ridges (divergent boundaries) produce basaltic oceanic crust. Subduction zones (convergent boundaries) generate a wider range of igneous rocks, from basalt to andesite to granite, through processes like partial melting and magmatic differentiation. Transform boundaries primarily deform existing rocks rather than creating new ones.
- Magmatic processes such as partial melting, fractional crystallization, and magma mixing determine the specific composition of igneous rocks. For example, partial melting of mantle peridotite at a mid-ocean ridge produces basaltic magma because lower-melting-point minerals melt first.
- Regional geological history matters too. A region that has experienced multiple tectonic events, burial and uplift, or repeated metamorphic episodes will have a more complex rock record than a stable continental interior (craton) that has remained relatively undisturbed for billions of years.

1.3 Plate tectonics and geodynamics

Plate tectonics is the unifying theory that explains large-scale motion and deformation of Earth's lithosphere. It connects earthquakes, volcanoes, mountain building, and seafloor spreading into a single coherent framework. Understanding how and why plates move is foundational to nearly everything else in geophysics, from hazard prediction to resource exploration to reconstructing Earth's history.

Plate Tectonics Theory

Key Concepts

The lithosphere is divided into several rigid plates that move relative to each other over the weaker, partially molten asthenosphere beneath them. Where these plates meet, you get plate boundaries, and the type of interaction at each boundary determines what geological processes occur there.

- Seafloor spreading: New oceanic crust forms at mid-ocean ridges as plates diverge. The seafloor moves away from the ridge axis, and continents drift apart over time.
- Subduction: At convergent boundaries, one plate sinks beneath another into the mantle. This recycles oceanic crust, generates volcanic arcs, and produces some of the planet's most powerful earthquakes.
- Transform faulting: Plates slide horizontally past each other along transform faults, producing significant lateral displacement and seismic activity.

Importance and Implications

Plate tectonics explains the global distribution of earthquakes, volcanoes, and mountain ranges. It also provides the basis for predicting where natural hazards like earthquakes, volcanic eruptions, and tsunamis are most likely to occur.

Beyond hazards, plate tectonics controls the formation and distribution of natural resources: mineral deposits, oil and gas reserves, and geothermal energy sources are all tied to tectonic processes. Over geological timescales, plate motion also influences the development of ecosystems and the distribution of species across continents.

Plate Boundaries and Features

Types of Plate Boundaries

Divergent boundaries form where two plates move apart. New oceanic crust is created at mid-ocean ridges through seafloor spreading. These boundaries are associated with rift valleys, volcanic activity, and shallow earthquakes.

- Examples: Mid-Atlantic Ridge, East Pacific Rise

Convergent boundaries form where two plates collide. The outcome depends on what type of lithosphere is involved:

- Oceanic-continental convergence: The denser oceanic plate subducts beneath the continental plate, producing deep-sea trenches and volcanic mountain ranges (e.g., the Andes).
- Oceanic-oceanic convergence: One oceanic plate subducts beneath the other, forming deep trenches and island arc volcanoes (e.g., the Mariana Trench).
- Continental-continental convergence: Neither plate subducts easily, so the crust crumples and thickens, building massive mountain ranges (e.g., the Himalayas).

All convergent boundaries are characterized by intense seismic activity.

Transform boundaries form where two plates slide horizontally past each other. These produce shallow to moderate earthquakes and often create linear valleys or ridges along the fault trace.

- Examples: San Andreas Fault (California), Alpine Fault (New Zealand)

Plate Boundary Zones

Not all plate boundaries are sharp, well-defined lines. Plate boundary zones are broad regions where the boundary between plates is diffuse, with complex deformation spread over a wide area.

- These zones often show a mix of divergent, convergent, and transform motion simultaneously.
- Examples include the Mediterranean region and the Himalayan-Tibetan orogen.
- Understanding them requires integrating seismic data, GPS velocity measurements, and geological field observations.

Driving Forces of Plate Motion

Mantle Convection

Mantle convection is the large-scale circulation of material within Earth's mantle, driven by heat transfer from the interior. Hot, buoyant material rises from the lower mantle while cold, dense material sinks from the upper mantle, creating a slow circulating flow that interacts with the overlying plates.

The thermal energy powering this convection comes from two main sources: radioactive decay of elements like uranium, thorium, and potassium, and residual heat left over from Earth's formation. Convection patterns can be modified by variations in mantle composition, mineral phase changes, and the insulating effect of continental lithosphere.

Ridge Push and Slab Pull

Two specific forces act directly on the plates themselves:

1. Ridge push: Mid-ocean ridges stand elevated above the surrounding ocean floor. The gravitational potential energy difference between the high ridge and the lower-lying plate pushes the plate away from the ridge.
2. Slab pull: At subduction zones, the cold, dense subducting slab sinks into the mantle under its own weight. This gravitational pull drags the rest of the attached plate toward the trench.

The combination of ridge push and slab pull is thought to be the dominant driver of plate motion, with mantle convection playing a more passive, facilitating role. The relative contribution of each force varies depending on factors like the age and density of the subducting slab and the length of the subduction zone. Older oceanic lithosphere is colder and denser, so slab pull tends to be stronger for plates with old subducting slabs.

Evidence for Plate Tectonics

Magnetic Anomalies

When new oceanic crust forms at mid-ocean ridges, magnetic minerals (primarily magnetite) in the cooling basalt align with Earth's magnetic field. Because Earth's magnetic field periodically reverses polarity, the seafloor records an alternating pattern of normal and reversed magnetic polarity stripes running parallel to the ridge axis.

These stripes are symmetric on either side of the ridge, which directly supports the concept of seafloor spreading. The pattern can be correlated with the independently established geomagnetic polarity timescale, allowing scientists to date the seafloor and calculate spreading rates.

Seismic Patterns

The global distribution of earthquakes closely traces plate boundaries, and the characteristics of those earthquakes reveal the type of plate interaction:

- Divergent boundaries: Shallow earthquakes along the ridge axis, caused by tensional (extensional) stress as plates pull apart.
- Convergent boundaries: Earthquakes range from shallow near the trench to deep along the subducting slab (forming a Wadati-Benioff zone), caused by compressional stress.
- Transform boundaries: Shallow to moderate earthquakes along the fault, caused by shear stress as plates slide past each other.

Focal mechanism solutions (also called "beach ball" diagrams) from earthquake seismology reveal the orientation of stress and the sense of slip on faults, providing direct information about plate motions and the forces acting at boundaries.

Age Progression of Seafloor and Volcanic Rocks

Radiometric dating and biostratigraphy show that seafloor age increases systematically with distance from mid-ocean ridges. The youngest crust is found at the ridge axis, and the oldest oceanic crust (about 200 million years old) is found near continental margins. No oceanic crust is older than ~200 Ma because older crust has been recycled back into the mantle through subduction.

Hotspot volcanism provides additional evidence. As a plate moves over a stationary mantle plume, it produces a chain of volcanic islands that get progressively older with distance from the hotspot. The Hawaiian-Emperor seamount chain is the classic example: the active volcanoes of Hawaii sit over the hotspot, while seamounts stretching northwest across the Pacific are progressively older, reaching ~80 Ma at the Emperor end.

Paleoclimatic Data

The distribution of ancient climate indicators on today's continents only makes sense if those continents have moved:

- Glacial deposits are found in regions that now sit at warm, equatorial latitudes. For example, glacial tillites in India and Australia date from when these landmasses were part of the southern supercontinent Gondwana, positioned near the South Pole.
- Coral reefs and limestone deposits occur in regions that are now at cold, high latitudes. Limestone in the Canadian Arctic formed when that region was near the equator.

These observations are fully consistent with the continental positions predicted by plate tectonic reconstructions.

Global Distribution and Geochemistry of Volcanic Rocks

The chemistry of volcanic rocks varies systematically with tectonic setting, providing independent confirmation of plate tectonic processes.

- Mid-ocean ridge basalts (MORB) erupt at divergent boundaries. They are depleted in incompatible elements (those that preferentially enter melts) and reflect the composition of the upper mantle source from which they are derived through decompression melting.

- Island arc volcanics erupt at convergent boundaries above subduction zones. They are enriched in incompatible elements and volatiles (water, CO_2), reflecting the addition of fluids and partial melts released from the subducting slab into the overlying mantle wedge.

These distinct geochemical signatures allow geologists to identify the tectonic setting of ancient volcanic rocks, even when the original plate boundary is no longer active.

Plate Tectonics and Geological Events

Earthquakes

Earthquakes are primarily concentrated at plate boundaries, where the interaction between plates causes elastic strain energy to build up in the lithosphere. When the accumulated stress exceeds the frictional strength of a fault, the rock ruptures and releases energy as seismic waves.

The type and depth of earthquakes vary by boundary type:

- Shallow earthquakes at divergent boundaries
- Shallow to deep earthquakes at convergent boundaries (the deepest reaching ~700 km along subducting slabs)
- Shallow to moderate earthquakes at transform boundaries

Studying earthquake patterns and focal mechanisms helps scientists characterize the regional stress field and plate motions, which is essential for seismic hazard assessment.

Volcanic Eruptions

Volcanic eruptions are most commonly associated with convergent boundaries, where volatiles released from the subducting slab lower the melting point of the overlying mantle wedge. The resulting magma rises to the surface and forms volcanic arcs (e.g., the Andes Volcanic Belt, Cascade Volcanic Arc).

Volcanism also occurs at divergent boundaries and hotspots. At divergent boundaries, the upwelling of hot mantle material undergoes decompression melting, producing basaltic magma (e.g., Iceland). At hotspots, a deep mantle plume generates volcanism independent of plate boundaries (e.g., the Hawaiian Islands).

Volcanic eruptions can have far-reaching impacts: ash and gas emissions affect air quality, climate, and aviation, while lava flows and pyroclastic density currents can devastate nearby communities and infrastructure.

Mountain Building (Orogenesis)

Mountain building is primarily driven by plate convergence. The two main mechanisms are:

1. Continental collision: When two continental plates converge, neither subducts easily due to their buoyancy. Instead, the crust compresses and thickens dramatically, producing extensive mountain ranges like the Himalayas and the Alps.

2. Subduction-related orogeny: Where oceanic crust subducts beneath continental crust, the associated magmatism, accretion of sediments, and crustal shortening build mountain ranges like the Andes.

The long-term evolution of mountain belts involves a feedback loop between tectonics and surface processes. Weathering and erosion remove material from the peaks, while isostatic adjustment causes the crust to rebound upward in response to the unloading. Sedimentation in adjacent basins adds load that drives subsidence. This interplay between tectonic uplift, erosion, and isostasy shapes the geomorphology of mountain ranges over millions of years.

Studying the tectonic history of mountain belts helps reconstruct past plate configurations, paleoclimate, and the coupled evolution of Earth's surface and interior.

Unit 2 – Seismology

2.1 Seismic waves and their properties

Seismic waves are the primary tool geophysicists use to probe Earth's interior. Because we can't directly sample most of what lies beneath the surface, we rely on how these waves travel, reflect, and attenuate to reconstruct the planet's layered structure.

Two broad categories exist: body waves that propagate through Earth's interior, and surface waves that travel along or near the surface. Their distinct behaviors in different materials form the basis of nearly everything we know about Earth's deep structure.

Seismic Wave Types and Characteristics

Body Waves

P-waves (primary or compressional waves) are the fastest seismic waves and arrive first on a seismogram. They propagate through both solids and liquids.

- Particle motion is parallel to the direction of propagation, producing alternating zones of compression and rarefaction.
- Velocities range from ~1.5 km/s in unconsolidated sediments to over 13 km/s in the inner core.

S-waves (secondary or shear waves) are slower and travel only through solid materials.

- Particle motion is perpendicular to the propagation direction, producing a shearing deformation.
- S-wave velocity is typically about 60% of the P-wave velocity in the same solid material.
- S-waves cannot propagate through liquids because liquids have zero shear modulus. This is why S-waves vanish at the outer core boundary, which is one of the strongest pieces of evidence that the outer core is liquid.

Surface Waves

Surface waves are generated when body waves interact with Earth's free surface. They carry most of the destructive energy during earthquakes.

Rayleigh waves produce elliptical particle motion in the vertical plane (both vertical and horizontal components), creating a ground-rolling effect.

- They are the slowest seismic waves and typically cause the most structural damage.
- Their velocity is slightly less than the S-wave velocity in the same material.

Love waves produce horizontal particle motion perpendicular to the propagation direction (side-to-side shearing).

- Faster than Rayleigh waves, but slower than body waves.
- Confined to shallow depths and require a low-velocity layer overlying a higher-velocity layer to propagate (a waveguide effect). Without this velocity contrast, Love waves don't exist.

Factors Influencing Seismic Wave Propagation

Elastic Properties and Density

Seismic wave velocity depends on the balance between a material's stiffness (elastic moduli) and its density. Stiffer materials transmit waves faster; denser materials slow them down.

The governing equations are:

- P-wave velocity: $V_p = \sqrt{\frac{K + \frac{4}{3}\mu}{\rho}}$
- S-wave velocity: $V_s = \sqrt{\frac{\mu}{\rho}}$

where K is the bulk modulus (resistance to uniform compression), μ is the shear modulus (resistance to shearing), and ρ is density.

Notice that V_s depends only on μ. When $\mu = 0$ (as in a liquid), $V_s = 0$, confirming that S-waves can't travel through fluids.

Temperature and pressure both change with depth, and they compete:

- Increasing temperature reduces elastic moduli by weakening atomic bonds and promoting thermal expansion. This tends to decrease velocity.
- Increasing pressure increases elastic moduli and density by compressing the material and closing pore space. This tends to increase velocity.

In most of Earth's interior, the pressure effect wins, so velocity generally increases with depth.

Anisotropy and Heterogeneity

Anisotropy means the elastic properties of a material vary with direction. Seismic waves traveling in different directions through the same rock will have different velocities.

- Common in layered sedimentary sequences and foliated metamorphic rocks, where mineral grains or layers create a preferred orientation.
- One observable consequence is shear wave splitting (birefringence): an S-wave entering an anisotropic region splits into two orthogonally polarized components that travel at different speeds.

Heterogeneities (inclusions, fractures, compositional changes) scatter and attenuate seismic energy.

- Scattering redistributes wave energy in multiple directions and generates coda waves, the long tail of scattered arrivals that follow the main phases on a seismogram.
- Attenuation from scattering preferentially removes high-frequency energy, because shorter wavelengths interact more strongly with small-scale heterogeneities.

Seismic Wave Velocity and Attenuation

Velocity Variations in Earth Materials

Seismic velocities span a wide range depending on material type and depth. Representative values:

Material	P-wave velocity (km/s)	S-wave velocity (km/s)
Unconsolidated sediments	1.5–2.5	0.2–1.0

Material	P-wave velocity (km/s)	S-wave velocity (km/s)
Consolidated sedimentary rocks	2.5–6.0	1.5–3.5
Crystalline rocks (igneous/metamorphic)	5.5–7.5	3.0–4.5
Mantle	7.5–13.0	4.5–7.5
Outer core (liquid)	8.0–10.5	—
Inner core (solid)	11.0–13.0	3.5–4.5

Velocity generally increases with depth, but not smoothly. Sharp jumps occur at major compositional or phase boundaries:

• Crustal velocity structure: gradual increase with depth, punctuated by the Moho at the base of the crust.

• Mantle velocity structure: gradual increase with depth, with notable discontinuities at ~410 km and ~660 km corresponding to mineral phase transitions (e.g., olivine to wadsleyite, ringwoodite to bridgmanite + ferropericlase).

Attenuation Mechanisms and Quality Factor

As seismic waves propagate, they lose energy. The quality factor Q quantifies this: higher Q means less energy loss per oscillation cycle, and more efficient propagation.

$Q(\omega) = \frac{\omega}{2\alpha(\omega)V}$

where ω is angular frequency, α is the attenuation coefficient, and V is wave velocity.

Three main mechanisms contribute to amplitude decay:

1.

Intrinsic absorption converts elastic energy to heat through anelastic processes like grain boundary sliding and dislocation motion. It increases with temperature and is strongest in partially molten or fluid-rich regions (the asthenosphere, magma chambers).

2.

Scattering redistributes energy due to heterogeneities. Its strength depends on the size and velocity contrast of the heterogeneities relative to the seismic wavelength. Highly fractured fault zones and volcanic regions show pronounced scattering.

3.

Geometrical spreading reduces energy density as the wavefront expands away from the source. This occurs even in a perfectly elastic medium: - Body wave amplitude decays as $1/r$ - Surface wave amplitude decays as $1/\sqrt{r}$

where r is distance from the source. The slower decay of surface waves is why they dominate seismograms at large epicentral distances.

Earth's Internal Structure from Seismic Waves

Seismic Wave Arrival Times and Amplitudes

The time difference between P-wave and S-wave arrivals (S-P time) at a station increases with distance from the earthquake source, because the two wave types travel at different speeds and the gap between them grows over longer paths.

- With S-P times from at least three stations, you can triangulate the epicenter using travel-time curves.
- Abrupt changes in the S-P time relationship at certain distances signal major internal discontinuities (e.g., the core-mantle boundary).

Amplitude information complements travel times. Lower-than-expected amplitudes along a particular path suggest the waves crossed a region of high attenuation, potentially indicating partial melt or elevated fluid content.

Seismic Discontinuities and Velocity Structure

At a discontinuity, seismic waves reflect and convert between P and S types. The major discontinuities that define Earth's layered structure:

- Mohorovičić discontinuity (Moho): crust-mantle boundary. P-wave velocity jumps from ~6.0–7.5 km/s to ~7.5–8.5 km/s. Depth varies from ~5–10 km under oceans to ~30–70 km under continents.
- Core-mantle boundary (CMB) at ~2,891 km depth: P-wave velocity drops sharply from ~13.0 km/s to ~8.0 km/s, and S-waves disappear entirely, indicating the outer core is liquid.
- Inner core boundary (ICB) at ~5,150 km depth: P-wave velocity increases from ~10.5 km/s to ~11.0 km/s, and S-waves reappear, indicating the inner core is solid.

Because velocity generally increases with depth, seismic rays curve back toward the surface according to Snell's law:

$$\frac{\sin\theta_1}{V_1} = \frac{\sin\theta_2}{V_2}$$

where θ is the angle of incidence or refraction and V is wave velocity. This curving creates shadow zones, angular ranges where direct P or S arrivals are absent. The P-wave shadow zone between ~104° and ~140° from the epicenter, and the complete absence of direct S-waves beyond ~104°, were the key observations that revealed the liquid outer core.

Seismic Tomography and 3D Earth Structure

Seismic tomography extends the 1D radial velocity model into three dimensions. The basic idea: if a wave arrives earlier than predicted by the 1D model, it traveled through faster-than-average material; if it arrives late, it crossed slower-than-average material.

Three main approaches:

1. Travel-time tomography: inverts residuals (observed minus predicted travel times) for 3D velocity perturbations. The most widely used method.
2. Attenuation tomography: inverts amplitude decay patterns for 3D variations in Q.
3. Waveform tomography: fits the full waveform (shape and amplitude), yielding higher-resolution models but at greater computational cost.

Tomographic images have revealed major lateral structures in the mantle:

- Mantle plumes: narrow, low-velocity, high-attenuation columns rising from near the CMB to the surface, associated with hotspot volcanism (e.g., Hawaii, Iceland).
- Subducting slabs: high-velocity, low-attenuation sheets extending from trenches down into the lower mantle, representing cold oceanic lithosphere sinking at convergent boundaries (e.g., beneath the Pacific Ring of Fire).
- Large low-shear-velocity provinces (LLSVPs): two broad, low-velocity regions in the lowermost mantle beneath Africa and the Pacific. Their origin remains debated, with hypotheses ranging from purely thermal anomalies to compositionally distinct material, possibly primordial.

2.2 Seismic instrumentation and data acquisition

Seismometer Components and Functioning

Seismometers are the foundation of observational seismology. They translate ground motion into electrical signals that can be recorded, stored, and analyzed. Understanding how they work, and where they fall short, is essential for interpreting any seismic dataset.

Seismometer Design and Components

A seismometer has three core components:

1. Sensor — detects ground motion. This is either a mass-spring system (a mass suspended on springs that stays relatively still while the frame moves with the ground) or a force-balance system (uses a feedback loop to actively hold the mass stationary relative to the ground, then measures the force required to do so).
2. Damping mechanism — suppresses unwanted oscillations near the instrument's resonant frequency. Common approaches include oil damping and electromagnetic damping. Without adequate damping, the sensor would "ring" after an impulse and distort the recorded waveform.
3. Transducer — converts the mechanical motion (or the feedback signal in force-balance designs) into an electrical signal for recording.

Force-balance sensors are standard in modern broadband instruments because the feedback design gives them a much flatter and wider frequency response than a purely passive mass-spring system.

Broadband Seismometers and Their Advantages

Traditional short-period seismometers are tuned to a narrow frequency band (typically around 1 Hz). Broadband seismometers overcome this limitation by recording across a wide frequency range, from long-period surface waves (periods of hundreds of seconds) down to short-period P and S waves (frequencies of tens of Hz).

Why this matters:

- They can detect smaller ground motions thanks to higher sensitivity across the spectrum.
- A single instrument can capture both local microseismic events and large teleseismic earthquakes (events thousands of kilometers away).
- Waveform analysis in both the time domain and frequency domain becomes possible from one record, giving you information about Earth structure and source properties simultaneously.

Seismic Arrays and Their Applications

A seismic array is a network of seismometers deployed in a deliberate geometric pattern. The primary goal is to improve the signal-to-noise ratio (SNR) beyond what any single station can achieve.

Common array geometries:

- Linear arrays — best for studying wave propagation along a specific azimuth.
- Circular arrays — provide omnidirectional sensitivity.
- Grid (2D) arrays — offer high spatial resolution for imaging 3D subsurface structure.

Key array processing techniques:

- Beamforming — signals from all stations are summed with calculated time delays so that energy arriving from a target direction adds constructively, while noise from other directions cancels out. Think of it as "steering" the array's sensitivity toward a particular source.
- Frequency-wavenumber (f-k) analysis — transforms the data into the frequency-wavenumber domain, where different seismic phases separate by their apparent velocity and propagation direction. This is especially useful for distinguishing overlapping arrivals that look similar in the time domain.

Digital Seismic Data Acquisition

Nearly all modern seismic recording is digital. The analog electrical signal from the seismometer is sampled, quantized, and stored as discrete numerical values. The quality of this digitization process directly controls what you can and cannot resolve in the data.

Analog-to-Digital Conversion and Sampling

An analog-to-digital converter (ADC) samples the continuous sensor output at regular time intervals.

The Nyquist-Shannon sampling theorem sets the fundamental rule: the sampling rate must be at least twice the highest frequency you want to record. If you violate this, you get aliasing, where high-frequency energy folds back and masquerades as lower-frequency signal. There's no way to fix aliased data after the fact.

For example, if your highest frequency of interest is 50 Hz, you need a sampling rate of at least 100 samples per second (100 Hz). In practice, oversampling (sampling well above the Nyquist rate) is common because it improves SNR and reduces the effects of quantization noise.

Dynamic Range and Seismic Data Formats

Dynamic range is the ratio between the largest and smallest amplitudes the system can faithfully record, expressed in decibels (dB) or bits.

- Modern systems typically use 24-bit ADCs, which provide roughly 144 dB of dynamic range. This is enough to record both the faint background hum of the Earth and the strong near-field signal of a moderate earthquake without clipping or losing small-amplitude detail.

Standard data formats:

- SEED (Standard for the Exchange of Earthquake Data) and its lighter variant miniSEED are the community standards. They bundle waveform data with metadata (station coordinates, instrument response, timing).

- Steim compression is a lossless algorithm commonly applied to miniSEED data. It exploits the sample-to-sample similarity in seismic traces to achieve high compression ratios without discarding any information.

These standardized formats are what make it possible for researchers worldwide to share and archive data through repositories like IRIS.

Seismic Data Processing Techniques

Raw seismic records contain the signal you want plus instrument effects, Earth attenuation, and noise. Processing strips away the unwanted parts and enhances the useful information.

Filtering and Deconvolution

Filtering selectively passes or rejects certain frequency bands:

Filter Type	What It Does
Low-pass	Removes high-frequency noise
High-pass	Removes low-frequency noise
Band-pass	Isolates a specific frequency range
Band-reject (notch)	Removes a specific frequency range

Filters can be applied in the time domain (via convolution) or in the frequency domain (via Fourier transforms). The choice often depends on computational convenience and the filter design.

Deconvolution goes a step further. Every seismometer has an instrument response, a transfer function describing how it converts true ground motion into voltage. Deconvolution removes this response (and optionally corrects for Earth attenuation) to recover the actual ground displacement, velocity, or acceleration. It can also compress the seismic wavelet, improving temporal resolution so you can distinguish closely spaced arrivals or reflectors.

The instrument response is determined through calibration experiments and is stored in the metadata alongside the waveform data.

Advanced Signal Analysis Techniques

Spectral analysis using the Fourier transform decomposes a time-domain signal into its constituent frequencies. This reveals the power spectrum (which frequencies carry the most energy) and phase information.

Polarization analysis applies to three-component recordings (vertical + two horizontal). It determines the direction of particle motion, which lets you separate wave types:

- P waves — linear particle motion parallel to the propagation direction.
- SV waves — particle motion perpendicular to propagation, in the vertical plane.
- SH waves — particle motion perpendicular to propagation, in the horizontal plane.

Seismic attribute analysis extracts derived quantities from the waveform to highlight subsurface features:

- Instantaneous amplitude — relates to signal energy; useful for spotting high-amplitude reflectors or lithology changes.
- Instantaneous phase — tracks wavefront continuity and reveals phase shifts or discontinuities.

• Instantaneous frequency — identifies changes in frequency content that may indicate variations in rock properties or the presence of fluids (including hydrocarbons in exploration contexts).

Seismic Instrumentation Limitations

Even the best instruments introduce artifacts. Recognizing and mitigating these limitations is just as important as understanding the instruments themselves.

Noise Sources and Their Mitigation

Instrument self-noise comes from electronic components and thermal fluctuations inside the seismometer. It sets a floor below which real ground motion cannot be detected. Low-noise instruments minimize this through high-quality electronics, electromagnetic shielding, and temperature-compensated circuitry.

Site noise comes from the environment:

• Wind — pressure fluctuations tilt and vibrate the sensor. Mitigation: wind shields, or array-based coherent noise cancellation.

• Ocean microseisms — generated by ocean wave interactions, these produce a persistent noise peak near 0.1–0.3 Hz. Mitigation: seafloor installations or pressure-sensor corrections for the water column.

• Cultural noise — traffic, machinery, footsteps. Mitigation: install stations in underground vaults, boreholes, or remote locations away from human activity.

Proper site selection is often the single most effective way to reduce noise.

Timing Errors and Quantization Noise

Clock drift in the data acquisition system causes timing errors that propagate directly into event location estimates. GPS synchronization and regular clock corrections keep drift to microsecond levels.

Quantization noise arises because the ADC rounds continuous voltage values to the nearest discrete level. Higher-resolution ADCs (24-bit vs. 16-bit) reduce this noise substantially. An additional technique called dithering adds a tiny amount of random noise to the analog signal before digitization. This sounds counterintuitive, but it randomizes the quantization error and effectively improves the ADC's resolution for small signals.

Metadata Management and Quality Control

Even perfect waveform data becomes useless if the metadata is wrong. Incorrect instrument response parameters, station coordinates, or timing information will propagate errors through every downstream analysis step.

Best practices for metadata and quality control:

• Review and update metadata whenever instrumentation or station configuration changes.

• Use automated QC tools to flag gaps, spikes, abnormal amplitudes, or timing jumps in continuous data streams.

• Supplement automated checks with visual inspection of waveforms and spectrograms to catch subtle artifacts that algorithms may miss.

Rigorous metadata management isn't glamorous, but it's what separates reliable seismic datasets from unreliable ones.

2.3 Seismic reflection and refraction methods

Seismic reflection and refraction methods are the primary tools geophysicists use to image Earth's subsurface. Both techniques generate seismic waves and record how those waves interact with underground layers, but they exploit different physical phenomena and serve different purposes. Reflection methods capture waves that bounce off subsurface interfaces, producing detailed structural images. Refraction methods capture waves that travel along interfaces, revealing large-scale velocity structure. Understanding how each works, when to use it, and what can go wrong is essential for designing surveys and interpreting results.

Seismic Reflection vs Refraction

Principles and Methods

Seismic reflection measures the two-way travel time of waves that bounce off subsurface interfaces such as sedimentary boundaries and faults. The physical basis is acoustic impedance contrast. Acoustic impedance (Z) is the product of seismic velocity and density:

$$Z = \rho v$$

Whenever a wave crosses an interface where Z changes significantly, part of its energy reflects back to the surface. The stronger the contrast, the stronger the reflection.

Seismic refraction measures the travel times of waves that are critically refracted along subsurface interfaces. The physical basis is Snell's law, which relates the angles of incidence and refraction at a boundary:

$$\frac{\sin\theta_1}{v_1} = \frac{\sin\theta_2}{v_2}$$

Critical refraction occurs when the refracted ray travels exactly along the interface ($\theta_2 = 90°$). This happens at the critical angle θ_c, where $\sin\theta_c = v_1/v_2$. The refracted wave then propagates at the velocity of the faster, deeper layer and radiates energy back to the surface as it goes.

A key requirement for refraction to work: velocity must increase with depth. If a low-velocity layer sits beneath a high-velocity layer, the refracted arrivals from deeper interfaces can be hidden, creating a "hidden layer" problem.

Applications and Survey Design

Reflection and refraction surveys differ in their targets, resolution, and logistics:

Feature	Reflection	Refraction
Best for	Detailed imaging of complex structures (basins, reservoirs, faults)	Large-scale velocity structure and crustal layering
Receiver spacing	Dense (meters to tens of meters)	Can be wider (hundreds of meters to kilometers)
Frequency content	Higher frequencies for better resolution	Lower frequencies that propagate over longer distances
Offset range	Moderate (target-dependent)	Long offsets needed to record critically refracted arrivals
Processing focus	Deconvolution, stacking, migration	Travel time picking, intercept-time analysis, tomographic inversion

The choice between methods depends on target depth, the resolution you need, and available budget. Many projects use both: refraction for a velocity framework and reflection for structural detail.

Seismic Survey Design

Survey Planning and Parameter Selection

Designing a seismic survey starts with defining the geologic target. The target's depth, size, and structural complexity drive every parameter choice.

Source selection depends on the environment: - Land: Explosives (strong, impulsive signal) or vibroseis trucks (controllable, repeatable sweep signal) - Marine: Air guns towed behind a vessel

Receiver selection also depends on the environment: - Land: Geophones, which measure ground velocity - Marine: Hydrophones, which measure pressure changes in water

Key parameter trade-offs:

- Deeper targets require longer source-receiver offsets and lower source frequencies, since high frequencies attenuate more rapidly with distance.
- Higher resolution demands denser spatial sampling (closer receiver spacing) and higher frequencies.
- Fold coverage (the number of traces that share a common midpoint) controls the signal-to-noise ratio after stacking. Higher fold means better noise suppression but more field effort.

Recording parameters like sampling rate, record length, and anti-alias filters should be set to capture the full bandwidth of useful signal without aliasing or truncating reflections from the deepest target.

Survey Logistics and Quality Control

Land surveys face practical constraints that shape the design: - Rugged terrain may require helicopter-portable equipment or specialized vibroseis vehicles. Survey geometries might shift from regular grids to crooked-line 2D or sparse 3D layouts. - Environmental regulations can restrict source types (no explosives near wetlands, for example) or limit access during sensitive wildlife seasons.

Marine surveys tow long hydrophone streamers (sometimes 8+ km) behind a vessel, with air gun arrays firing at regular intervals: - Streamer length and the number of parallel streamers determine subsurface coverage and crossline resolution. - GPS and acoustic positioning systems track source and receiver locations precisely. Even small positioning errors degrade image quality.

Quality control during acquisition is critical. Field crews monitor source output (consistent signatures), receiver responses (no dead or noisy channels), and ambient noise levels in real time. Data gaps from equipment failures or access problems are much cheaper to fill during acquisition than to work around in processing.

Seismic Data Interpretation

Reflection Data Processing and Imaging

Raw seismic reflection data is noisy and geometrically distorted. Processing transforms it into an interpretable subsurface image. Here's the standard workflow:

1. Trace editing — Remove dead, noisy, or corrupted traces that would degrade the final image.

2. Amplitude recovery — Compensate for energy loss due to geometric spreading (amplitude decreases with distance) and absorption (the earth converts seismic energy to heat).

3. Deconvolution — Compress the source wavelet to sharpen temporal resolution. This removes reverberations and the imprint of the source signature.

4. Velocity analysis — Estimate how seismic velocity varies with depth by examining how reflection travel times change with offset. This step is iterative and directly affects image quality.

5. Normal moveout (NMO) correction — Reflections from a flat layer arrive later at far offsets than near offsets, forming a hyperbolic curve. NMO correction flattens these hyperbolas so that all traces for a given reflector align at the same time.

6. Stacking — Sum all NMO-corrected traces that share a common midpoint (CMP). This boosts the signal (reflections add constructively) and suppresses random noise.

7. Migration — Reposition reflectors to their true subsurface locations. Without migration, dipping reflectors appear shifted, and diffraction patterns from faults or pinch-outs remain unresolved. Migration also collapses the Fresnel zone, improving horizontal resolution.

For geologically complex areas (salt bodies, thrust belts), advanced techniques like pre-stack depth migration and full-waveform inversion can produce more accurate images by handling strong lateral velocity variations that simpler methods cannot.

Refraction Data Processing and Velocity Modeling

Refraction processing focuses on extracting velocity and layer-thickness information from first-arrival travel times.

1. Pick first arrivals on each trace. These are the earliest arriving energy, typically refracted waves at longer offsets and direct waves at short offsets.

2. Plot travel time vs. offset curves. Each straight-line segment on this plot corresponds to a different subsurface layer. The slope of each segment equals $1/v$, where v is the velocity of that layer. The intercept time (where the line extrapolates back to zero offset) relates to the depth and velocity of overlying layers.

3. Invert for layer model. Simple cases (flat, horizontal layers) can be solved with intercept-time formulas. More complex structures require tomographic inversion, which iteratively adjusts a velocity model until predicted travel times match observed ones. Two common approaches: - Ray tracing — Simulates individual ray paths through the model - Wavefront methods — Solve the eikonal equation to compute travel times across a gridded model

The resulting velocity model is most useful when integrated with other data. Well logs provide ground-truth velocities and lithologies at specific points. Gravity and magnetic data add constraints on density and composition. Seismic attributes extracted from reflection data (amplitude, phase, frequency variations) can further characterize reservoir properties like porosity, lithology, and fluid content.

Seismic Method Limitations

Resolution and Accuracy Constraints

Vertical resolution is governed by the dominant wavelength (λ) of the seismic signal:

$\lambda = \frac{v}{f}$

where v is velocity and f is dominant frequency. The commonly cited vertical resolution limit is approximately $\lambda/4$. For example, at $v = 3000$ m/s and $f = 50$ Hz, $\lambda = 60$ m, so vertical resolution is roughly 15 m. Layers thinner than this may still produce a reflection, but you can't resolve their top and bottom as separate interfaces.

Horizontal resolution before migration is limited by the Fresnel zone, the area on a reflector that contributes constructively to a single reflection event. The Fresnel zone radius grows with depth and wavelength. Migration processing collapses the Fresnel zone, sharpening horizontal resolution to roughly half the dominant wavelength under ideal conditions.

Velocity model accuracy depends on the quality of travel time picks, the assumptions built into inversion algorithms (e.g., isotropy, layer geometry), and how well the survey geometry samples the subsurface. Seismic anisotropy, caused by aligned fractures, fine layering, or mineral fabric, makes velocity direction-dependent. If anisotropy isn't accounted for, depth conversions and structural interpretations can contain significant errors.

Geologic Complexity and Interpretation Challenges

Certain geologic settings push seismic methods to their limits:

- Salt bodies have much higher velocities than surrounding sediments and often have irregular shapes. They distort the wavefield passing through them, creating shadow zones and imaging artifacts beneath the salt.
- Volcanic intrusions scatter seismic energy and introduce sharp velocity anomalies, making it difficult to image structures below them.
- Steeply dipping layers may not reflect energy back to the surface receivers (limited illumination), leaving gaps in the image.

Seismic data alone has limited sensitivity to fluid type and pore pressure. Amplitude variation with offset (AVO) analysis can provide some fluid and lithology discrimination by examining how reflection strength changes with offset, but it requires careful calibration. Well logs and core samples remain essential for tying seismic responses to actual rock and fluid properties.

Non-uniqueness is a fundamental challenge: multiple geologic models can produce the same seismic response. A bright reflection could be a gas sand, a coal seam, or a cemented layer. Interpreters constrain their models using regional geologic knowledge, well ties, and complementary geophysical datasets.

Finally, cost and environmental impact can limit where seismic surveys are feasible. Surveys may be restricted near populated areas, in protected habitats, or during sensitive ecological periods. Mitigation measures like low-energy sources, seasonal timing restrictions, and marine mammal observers are increasingly standard practice.

2.4 Earthquake seismology and seismic hazard assessment

Earthquake generation mechanisms

Earthquakes result from the sudden release of stored elastic strain energy in the Earth's crust and upper mantle, primarily along pre-existing fault zones. Understanding the mechanics behind this process is central to both seismology and hazard assessment.

Causes and processes of earthquake occurrence

Elastic rebound theory is the foundational model here. Tectonic forces slowly deform rocks on either side of a fault, storing elastic strain energy over years to centuries. When the accumulated stress exceeds the fault's frictional resistance, the fault ruptures and the two sides snap back toward their undeformed positions, releasing energy as seismic waves.

Several factors control whether and when an earthquake occurs:

- The local and regional stress field (driven by plate tectonics)
- The frictional properties and geometry of the fault surface
- The rate of stress accumulation from tectonic loading
- The presence of fluids in the crust, which can reduce frictional strength and promote failure

The tectonic setting determines how often and how large earthquakes can be. The largest and most frequent events occur along plate boundaries, particularly in subduction zones (e.g., Japan, Chile) and continental collision zones (e.g., the Himalayan region). Intraplate earthquakes are rarer but still significant.

Earthquake magnitude and energy release

Earthquake magnitude quantifies the energy released during rupture. It depends on three physical parameters: the rupture area, the average slip on the fault, and the rigidity (shear modulus) of the surrounding rock.

The moment magnitude scale (M_w) is the standard measure used today because it remains accurate across all earthquake sizes, unlike the older Richter scale (M_L), which saturates for large events. Moment magnitude is calculated from the seismic moment:

$$M_w = \frac{2}{3}\log_{10}(M_0) - 10.7$$

where M_0 is the seismic moment (in dyne·cm), defined as:

$$M_0 = \mu \cdot A \cdot D$$

Here μ is the shear modulus of the rock, A is the fault rupture area, and D is the average displacement (slip) across the fault.

Because the scale is logarithmic, energy increases dramatically with each unit of magnitude. A magnitude 7 earthquake releases roughly 32 times more energy than a magnitude 6, and about 1,000 times more than a magnitude 5.

Locating earthquake epicenters

Seismic waves and their propagation

Earthquakes generate seismic waves that travel through the Earth's interior and along its surface. Seismometers at stations around the world record these arrivals, and the data are used to pinpoint where the earthquake occurred.

The two main body wave types are:

- P-waves (primary/compressional): The fastest seismic waves. They compress and expand material in the direction of propagation and can travel through solids, liquids, and gases.

- S-waves (secondary/shear): Slower than P-waves. They move material perpendicular to the direction of propagation and can only travel through solids (they cannot pass through the outer core).

Locating an epicenter relies on the difference in P- and S-wave arrival times at multiple stations:

1. Record the arrival times of P- and S-waves at each station.

2. Calculate the S-P time interval (Δt) at each station. A larger Δt means the station is farther from the epicenter.

3. Use travel-time curves to convert each Δt into a distance estimate.

4. Draw a circle of that radius around each station on a map. With three or more stations, the circles intersect at (or near) the epicenter. This is the triangulation method.

The depth of the earthquake focus (hypocenter) is determined simultaneously, often by comparing observed arrival times against theoretical travel-time models for different depths.

Focal mechanisms and beach ball diagrams

The focal mechanism describes the orientation of the fault plane and the direction of slip during rupture. It's determined by analyzing the first-motion polarities (compressions vs. dilatations) and amplitudes of P-waves recorded at stations surrounding the earthquake.

Beach ball diagrams are the standard graphical representation. They project the pattern of compressional (shaded) and dilatational (white) first motions onto a lower-hemisphere stereographic plot. The pattern directly tells you the style of faulting:

- Normal faulting (extensional): The shaded quadrants form a "T" shape, with the center filled. Common in rift zones (e.g., East African Rift).

- Reverse/thrust faulting (compressional): The shaded quadrants are on the sides, with the center white. Found in subduction zones and collision zones (e.g., Andes, Himalayas).

- Strike-slip faulting (lateral): Four alternating quadrants of shaded and white, forming an "X" pattern. Characteristic of transform boundaries (e.g., San Andreas Fault).

Each type corresponds to a specific stress regime, so focal mechanisms are a powerful tool for mapping the tectonic stress field of a region.

Seismic hazards and risk

Tectonic settings and their seismic hazard characteristics

Seismic hazard is the probability that a given level of ground shaking (or related effect like liquefaction or landslides) will occur at a specific location within a defined time period. It's a physical quantity, distinct from risk.

Different tectonic settings produce distinct hazard profiles:

- Subduction zones generate the world's largest earthquakes ($M_w > 9$), such as the 2011 Tohoku event (M_w 9.1). They also pose major tsunami hazards to coastal communities because large seafloor displacements can displace enormous volumes of water.

- Continental collision zones (e.g., the Himalayan front) produce frequent moderate-to-large earthquakes. Seismic risk here is compounded by dense populations and vulnerable building stock, as seen in the 2015 Nepal earthquake (M_w 7.8).

• Intraplate regions experience less frequent earthquakes, but these can still be damaging. Recurrence intervals are long and poorly constrained, making hazard assessment more uncertain. The 1811-1812 New Madrid earthquakes in the central United States are a classic example.

Factors influencing seismic risk

Seismic risk combines hazard with exposure and vulnerability. A large earthquake in an uninhabited desert poses hazard but little risk; a moderate earthquake beneath a densely built city can be catastrophic.

Geographic and site-specific factors that amplify hazard:

• Proximity to active faults: Closer sites experience stronger shaking.

• Site conditions: Soft soils (e.g., alluvial basins, reclaimed land) can amplify seismic waves significantly compared to hard bedrock sites. This effect was dramatically demonstrated in the 1985 Mexico City earthquake, where lake-bed sediments amplified shaking far beyond what bedrock sites experienced.

• Topographic effects: Ridges and steep slopes can focus and amplify ground motions, and they increase the risk of earthquake-triggered landslides.

Vulnerability depends on the built environment:

• Construction type, age, design standards, and maintenance all matter. Unreinforced masonry buildings, for instance, are far more susceptible to collapse than modern steel-frame or reinforced concrete structures designed to seismic codes.

• Older structures built before modern seismic codes were adopted are consistently among the most vulnerable.

Seismic hazard assessment and mitigation

Probabilistic and deterministic hazard analysis methods

Two main frameworks exist for formal hazard assessment, and they serve different purposes.

Probabilistic Seismic Hazard Analysis (PSHA) is the more widely used approach. It quantifies the probability of exceeding a given ground motion level at a site over a specified time period (e.g., 10% probability of exceedance in 50 years). The basic steps are:

1. Identify and characterize seismic sources (known faults, areal source zones) in the region.

2. Define recurrence relationships for each source, specifying how often earthquakes of various magnitudes occur (typically using the Gutenberg-Richter relation or characteristic earthquake models).

3. Select ground motion prediction equations (GMPEs) that estimate shaking intensity as a function of magnitude, distance, and site conditions.

4. Integrate over all sources, magnitudes, and distances to produce hazard curves (annual probability of exceedance vs. ground motion level) and hazard maps.

PSHA accounts for uncertainty in earthquake location, size, and timing, making it well-suited for building codes and regional planning.

Deterministic Seismic Hazard Analysis (DSHA) takes a different approach: it identifies the largest credible earthquake on a specific fault or source zone and estimates the resulting ground motion at the site. There's no probability attached; it's a worst-case scenario assessment. DSHA is typically used for critical facilities like nuclear power plants, large dams, and major bridges, where the consequences of failure are severe enough to justify a conservative design basis.

Risk reduction and mitigation strategies

Seismic hazard maps produced by PSHA or DSHA feed directly into practical risk reduction:

• Building codes and design standards specify minimum levels of earthquake resistance for new construction. Implementing and enforcing these codes is the single most effective way to reduce earthquake casualties.

• Retrofitting existing structures (e.g., adding steel bracing to unreinforced masonry, base isolation for critical buildings) addresses the large stock of vulnerable older buildings.

• Microzonation studies map local variations in hazard due to soil conditions, topography, and proximity to faults. These guide land-use planning decisions, steering development away from the highest-risk zones or requiring enhanced design measures.

• Earthquake early warning (EEW) systems detect P-waves near the source and transmit alerts to more distant locations before the damaging S-waves and surface waves arrive. Even a few seconds of warning allows automated responses (stopping trains, opening fire station doors, triggering shutoffs for gas lines).

• Emergency response planning and regular drills ensure that communities can respond effectively when an earthquake strikes.

• Public education about seismic hazards, drop-cover-hold-on procedures, and household preparedness (securing heavy furniture, maintaining emergency supplies) builds long-term resilience.

Unit 3 – Gravity and Geodesy

3.1 Gravity and its measurement

Gravity: Concept and Principles

Gravity governs the large-scale structure of the Earth and the motion of objects within and around it. In geophysics, precise gravity measurements reveal density variations beneath the surface, making gravity one of the primary tools for mapping subsurface geology. This section covers the physical principles behind gravity, the instruments used to measure it, and how geophysicists interpret gravity data to identify hidden geological structures.

Fundamental Force of Nature

Gravity is the attractive force between any two objects that have mass. It's one of the four fundamental forces of nature, alongside the electromagnetic force, the strong nuclear force, and the weak nuclear force. Of these four, gravity is by far the weakest, but it dominates at planetary and astronomical scales because it acts over unlimited distances and is always attractive.

Newton's Law of Universal Gravitation

Newton's law gives us the quantitative framework for gravity:

$F = G\frac{m_1 m_2}{r^2}$

where F is the gravitational force, m_1 and m_2 are the masses of the two objects, r is the distance between their centers, and G is the universal gravitational constant (6.674×10^{-11} N·m^2/kg^2).

Two things control the strength of the force:

- Mass: Greater mass means a stronger pull. This is why density contrasts underground produce measurable gravity variations at the surface.
- Distance: The force drops off as $1/r^2$. Double the distance, and the force falls to one-quarter.

For geophysics, we're usually interested in the gravitational acceleration at Earth's surface, g, which averages about $9.81\ \text{m/s}^2$. Small spatial variations in g tell us about differences in rock density below.

Gravity's Role in the Universe

- Keeps planets in orbit around the Sun and the Moon in orbit around Earth
- Drives the formation of large-scale structures like galaxies and galaxy clusters
- Controls isostatic equilibrium in Earth's crust, which directly matters for geodesy

Einstein's Theory of General Relativity

General relativity describes gravity not as a force but as the curvature of spacetime caused by mass and energy. For most geophysical applications, Newtonian gravity is sufficient. General relativity becomes necessary in extreme conditions (near black holes, in the early universe) and for high-precision satellite geodesy, where relativistic corrections affect clock rates on orbiting instruments like those in GPS and GRACE.

Measuring Gravity

Gravimeters

A gravimeter measures the strength of the gravitational field at a specific location. Gravimeters fall into two broad categories: absolute and relative. The choice between them depends on the required accuracy, portability, and survey design.

Absolute Gravimeters

Absolute gravimeters measure the actual value of g at a location by directly tracking the acceleration of a free-falling object in a vacuum.

How they work (step by step):

1. A test mass (typically a corner-cube mirror) is released inside an evacuated chamber.
2. A laser interferometer tracks the position of the falling mass at precisely timed intervals.
3. The acceleration is computed from the position-time data using $s = \frac{1}{2}gt^2$.
4. Multiple drops (often hundreds) are averaged to reduce noise.

Common instruments include the FG5 (lab-grade, accuracy ~1–2 μGal) and the A10 (field-portable, accuracy ~10 μGal). These are expensive, heavy, and require a stable environment, so they're typically used to establish reference stations rather than for broad area surveys.

Relative Gravimeters

Relative gravimeters measure the difference in g between two locations rather than the absolute value. They're more portable and far less expensive than absolute instruments, making them the workhorse of field gravity surveys.

Two main types:

- Spring-based gravimeters (e.g., LaCoste & Romberg, Scintrex CG-5): A mass on a spring stretches or compresses in response to changes in gravity. The displacement is proportional to the change in g. These instruments are subject to drift over time and must be regularly returned to a base station for recalibration.
- Superconducting gravimeters: A niobium sphere is levitated by a magnetic field generated by superconducting coils. Changes in the current needed to maintain levitation correspond to changes in gravity. These are extremely sensitive (sub-μGal precision) but are stationary instruments used for continuous monitoring, not field surveys.

Gravity Gradiometers

A gravity gradiometer measures the spatial rate of change (gradient) of the gravitational field, not just its magnitude. This is the second derivative of the gravitational potential, expressed in units of Eötvös ($1\,\text{E} = 10^{-9}\,\text{s}^{-2}$).

Gradiometers are particularly useful because they emphasize shallow, localized density contrasts. They're widely used in airborne surveys for mineral and oil exploration, where they can detect features that standard gravimeters would miss.

Satellite-Based Methods

The GRACE (Gravity Recovery and Climate Experiment) mission, and its successor GRACE-FO, measure variations in Earth's gravity field from orbit. The technique works like this:

1. Two identical satellites orbit Earth about 220 km apart.
2. A microwave ranging system continuously measures the distance between them to micrometer precision.
3. As the lead satellite passes over a mass anomaly (e.g., a region of thicker crust), it accelerates slightly, changing the inter-satellite distance.
4. These distance changes are inverted to produce monthly maps of Earth's gravity field.

GRACE data have been used to track ice sheet mass loss, groundwater depletion, and post-glacial rebound, in addition to mapping the static gravity field.

Interpreting Gravity Data

Gravity Anomalies

A gravity anomaly is the difference between the observed value of g at a location and the theoretical value predicted by a reference model (which accounts for latitude, elevation, and other known factors). Anomalies are typically reported in milligals (mGal).

- A positive anomaly means observed gravity exceeds the predicted value, indicating denser-than-expected material below the surface (e.g., a mafic intrusion).
- A negative anomaly means observed gravity is lower than predicted, suggesting less dense material (e.g., a sedimentary basin or salt dome).

Bouguer Anomalies

The Bouguer anomaly is the gravity anomaly after applying several corrections to remove known, non-geological effects:

1. Free-air correction: Accounts for the station's elevation above the reference ellipsoid.
2. Bouguer correction: Removes the gravitational effect of the rock mass between the station and the reference ellipsoid, assuming a uniform slab of known density (typically $2670\,\text{kg/m}^3$ for average crustal rock).
3. Terrain correction: Compensates for the gravitational effect of nearby topographic irregularities that the simple slab model doesn't capture.

The resulting Bouguer anomaly reflects subsurface density variations and is the primary quantity used to identify structures like sedimentary basins (negative Bouguer anomalies), igneous intrusions (positive), and ore bodies.

Isostatic Anomalies

An isostatic anomaly is the Bouguer anomaly with an additional correction for isostatic compensation. Isostasy is the principle that Earth's crust and upper mantle adjust vertically to maintain gravitational equilibrium, similar to blocks of different thickness floating in a fluid.

If the isostatic anomaly is near zero, the crust is in isostatic equilibrium. Significant non-zero values indicate regions where the crust has not fully adjusted, perhaps due to recent glacial unloading, tectonic loading, or active mantle dynamics. These anomalies provide insights into crustal thickness and upper mantle density.

Gravity Maps

Gravity maps display the spatial distribution of anomalies (usually Bouguer anomalies) across a survey area. They're used to:

- Identify geological structures such as faults, folds, and intrusions
- Guide mineral and hydrocarbon exploration by highlighting density contrasts
- Constrain models of crustal and lithospheric structure

Color-contoured gravity maps are often combined with other geophysical data (magnetics, seismic) for integrated interpretation.

Absolute vs. Relative Gravity Measurements

Absolute Gravity Measurements

- Determine the actual gravitational acceleration at a specific location
- Expressed in $\mathrm{m/s^2}$ or in Gal units, where $1\,\mathrm{Gal} = 1\,\mathrm{cm/s^2}$ (so $g \approx 981\,\mathrm{Gal}$)
- Independent of any reference point
- Require expensive, complex instrumentation (FG5, A10) and a stable operating environment
- Best suited for establishing base stations and calibrating gravity networks

Relative Gravity Measurements

- Determine the difference in gravitational acceleration between locations
- Expressed in mGal ($1\,\mathrm{mGal} = 10^{-3}\,\mathrm{Gal} = 10^{-5}\,\mathrm{m/s^2}$)
- Require a reference point (base station) with a known absolute gravity value
- Use more portable, less expensive instruments (spring gravimeters, superconducting gravimeters)
- More efficient for covering large survey areas in the field

Gravity Network Adjustment

To produce consistent, accurate gravity maps, relative measurements must be tied to an absolute reference frame. This is done through network adjustment:

1. Absolute gravity values are established at a set of base stations using absolute gravimeters.
2. Relative gravimeters are used to measure differences between these base stations and survey points.
3. A least-squares adjustment distributes measurement errors across the network, producing a self-consistent set of gravity values for all stations.

This process ensures that surveys conducted at different times or by different teams can be combined into a single, reliable dataset.

3.2 Gravity anomalies and their interpretation

Gravity anomalies in geophysics

Gravity anomalies reveal what's hidden beneath Earth's surface by measuring how the actual gravitational pull at a location differs from what we'd theoretically expect. These differences arise from lateral density variations in the subsurface, and interpreting them is one of the core skills in exploration geophysics and geodesy.

Definition and significance

A gravity anomaly is the difference between the observed gravitational acceleration at a point and the theoretical value predicted by a reference model (typically a reference ellipsoid). Lateral density contrasts within the Earth's interior drive these variations.

- A positive anomaly means gravity is stronger than expected, pointing to denser-than-average material below (e.g., a mafic intrusion).
- A negative anomaly means gravity is weaker than expected, pointing to less dense material (e.g., a thick sedimentary basin or salt body).

Gravity anomalies are central to geophysics because they provide direct information about subsurface density structure. That makes them useful for mapping geological features you can't see at the surface, from crustal thickness variations down to individual ore bodies.

Applications in geosciences

- Sedimentary basin delineation: Low-density sedimentary fill (shales, sandstones) produces broad negative anomalies, making gravity surveys a first-pass tool for oil and gas exploration.
- Subsurface structural mapping: Faults, folds, and unconformities create density contrasts that show up as gradients or offsets in gravity data. This helps reconstruct the tectonic history of a region.
- Mineral exploration: Dense ore bodies like iron ore, chromite, or massive sulfide deposits generate localized positive anomalies, guiding exploration targeting.
- Crustal and upper mantle studies: Regional gravity data constrain the depth to the Moho (crust-mantle boundary) and can indicate features like mantle upwellings or lithospheric downwellings.
- Integrated interpretation: Gravity data are routinely combined with magnetic anomalies, seismic reflection/refraction data, and borehole logs. This multi-method approach reduces the inherent non-uniqueness of potential field interpretation, where multiple subsurface models can produce the same surface anomaly.

Calculating gravity anomalies

Measurement and data acquisition

The basic idea: measure gravity at a location, then subtract what theory predicts for that location. The remainder is your anomaly.

• Observed gravity is measured with a gravimeter, an instrument that records the local acceleration due to gravity (g_{obs}) with precision on the order of 0.01 mGal or better.

• Theoretical (normal) gravity is computed from a reference ellipsoid using the International Gravity Formula or its modern equivalent (e.g., GRS80). This accounts for Earth's shape, mass distribution, and rotation, and it varies primarily with latitude:

$$g_{theoretical} = g_e \frac{1 + k\sin^2\phi}{\sqrt{1 - e^2\sin^2\phi}}$$

where ϕ is latitude, g_e is gravity at the equator, k is a derived constant, and e is the ellipsoid's eccentricity.

The raw anomaly is then:

$$\Delta g = g_{obs} - g_{theoretical}$$

But this raw difference isn't very useful yet. Several corrections must be applied first.

Corrections and processing

Each correction removes a known, non-geological effect so that the final anomaly reflects only subsurface density variations.

1. Free-air correction: Accounts for the station's elevation above the reference ellipsoid. Gravity decreases with height, so this correction adds approximately $0.3086\ \text{mGal/m}$ of elevation. The result is the free-air anomaly:

$$\Delta g_{FA} = g_{obs} - g_{theoretical} + 0.3086 \cdot h$$

where h is the station elevation in meters.

1. Bouguer correction: Removes the gravitational attraction of the rock mass between the station and the ellipsoid (modeled as an infinite horizontal slab). For a slab of density ρ:

$$\Delta g_{Bouguer\,slab} = 2\pi G\rho h$$

A standard crustal density of $2670\ \text{kg/m}^3$ is commonly used. Subtracting this from the free-air anomaly gives the simple Bouguer anomaly.

1.

Terrain correction: The Bouguer slab approximation assumes flat topography. Where hills rise above or valleys dip below the station, a terrain correction compensates for the actual shape of the surrounding landscape. Both hills and valleys cause the simple Bouguer anomaly to underestimate the true value, so terrain corrections are always positive.

2.

Complete Bouguer anomaly: After applying all three corrections, you get the complete Bouguer anomaly, which is the standard product for geological interpretation.

Gravity anomalies are expressed in milliGals (mGal), where $1\ \text{mGal} = 10^{-5}\ \text{m/s}^2$. Typical crustal anomalies range from a few mGal to several hundred mGal.

Interpreting gravity anomaly maps

Visualizing gravity anomalies

Gravity anomaly maps display the spatial distribution of anomaly values across a survey area, using either color scales or contour lines.

- Warm colors (reds) typically represent positive anomalies, suggesting relatively dense subsurface material such as igneous intrusions or uplifted basement rock.
- Cool colors (blues) represent negative anomalies, suggesting low-density features like sedimentary basins, salt domes, or voids.
- Steep gradients (closely spaced contours) often mark the edges of density contrasts, such as fault contacts or basin margins.

Analyzing gravity anomaly profiles

A gravity anomaly profile is a cross-section extracted along a line across the map. Profiles are powerful because the shape of the anomaly curve encodes information about the source body:

- Amplitude relates to the density contrast ($\Delta\rho$) and the size of the body. A large, dense intrusion produces a high-amplitude anomaly.
- Wavelength (the horizontal breadth of the anomaly) relates to the depth of the source. Deeper bodies produce broader, smoother anomalies; shallow bodies produce sharper, narrower ones. This is a direct consequence of potential field theory: the signal attenuates and spreads with distance.
- Shape and symmetry give clues about geometry. A symmetric bell-shaped anomaly suggests a roughly equidimensional body (like a sphere or vertical cylinder), while an asymmetric anomaly might indicate a dipping contact or fault.

Interpretation typically involves forward modeling (assuming a subsurface geometry, computing its predicted anomaly, and comparing to the observed data) or inversion (mathematically deriving a density model that fits the observed anomaly). Both approaches must be constrained by geological knowledge because gravity inversion is inherently non-unique.

Gravity anomalies for geological problems

Subsurface structure and tectonics

Gravity data are especially valuable for mapping large-scale crustal structure:

- Faults and basin margins appear as linear gravity gradients. The magnitude of the gradient relates to the throw and density contrast across the fault.
- Crustal thickness variations produce long-wavelength regional anomalies. Thicker crust (e.g., beneath mountain belts) tends to show negative Bouguer anomalies because of the deep, low-density crustal root, consistent with isostatic compensation.
- Moho depth can be estimated from gravity data using isostatic models or spectral analysis techniques. For example, the Bouguer anomaly over the Tibetan Plateau reaches roughly -500 mGal, reflecting a crustal root extending to ~70 km depth.
- Mantle dynamics: Broad positive anomalies over oceanic ridges or hotspots can indicate mantle upwellings, while negative anomalies in subduction zones may reflect the downgoing slab geometry.

Combining gravity with seismic and magnetic data produces more tightly constrained models and is standard practice in both academic research and industry.

Resource exploration

Gravity surveys remain a cost-effective reconnaissance tool for resource exploration:

- Oil and gas: Sedimentary basins filled with low-density shale and sandstone generate regional negative anomalies. Mapping basin geometry with gravity helps prioritize areas for more expensive seismic surveys. Salt domes, which trap hydrocarbons, also produce distinctive anomalies because salt ($\rho \approx 2160\ \mathrm{kg/m^3}$) is less dense than the surrounding compacted sediments.
- Mineral deposits: Dense ore bodies produce localized positive anomalies. For instance, a massive sulfide deposit with $\rho \approx 4000\text{–}4500\ \mathrm{kg/m^3}$ embedded in country rock at $\rho \approx 2700\ \mathrm{kg/m^3}$ creates a strong, compact anomaly that can be modeled to estimate the deposit's depth, size, and tonnage.
- Drill site selection: Gravity modeling helps optimize where to drill confirmation boreholes, reducing exploration costs by narrowing the target area before committing to expensive drilling programs.

3.3 Isostasy and its applications

Isostasy explains how Earth's crust floats on the denser mantle, balancing weight and buoyancy. This concept is central to understanding gravity anomalies, crustal thickness variations, and why the Earth's surface moves vertically over time.

Geophysicists use several models to describe isostasy, including the Airy, Pratt, and flexural approaches. These models help interpret gravity data, estimate crustal properties, and explain dynamic processes like mountain building and basin formation.

Isostasy: Concept and Principles

Isostatic Equilibrium and Compensation

Isostasy is the state of gravitational equilibrium between Earth's crust and mantle, where the lighter crust "floats" on the denser mantle much like an iceberg floats on water. Thicker or less dense crust rides higher, while thinner or denser crust sits lower.

Isostatic equilibrium is reached when the weight of a crustal column is exactly balanced by the buoyancy force from the mantle, so there's no net vertical force. The depth of compensation is the depth below which pressure from overlying columns becomes equal everywhere. This depth roughly corresponds to the base of the lithosphere, though it's often discussed in relation to the Moho (crust-mantle boundary), which varies with crustal thickness and density.

Isostatic Adjustments and Disturbances

When something disturbs the equilibrium, the crust slowly rises or sinks to restore balance. These are isostatic adjustments, and they're driven by processes that add or remove mass at the surface:

- Glaciation/deglaciation: Ice sheets load the crust downward; when they melt, the crust rebounds. Scandinavia is still rising at up to ~10 mm/yr after the last ice age, and Hudson Bay in Canada shows similar post-glacial rebound.
- Erosion and deposition: Removing material from mountains causes uplift, while depositing sediment in basins causes subsidence (e.g., the Gulf Coast and North Sea basins).

- Tectonic loading: Mountain building (Himalayas, Andes) or emplacement of dense igneous intrusions (Bushveld Complex, South Africa) can push the crust out of equilibrium.

The timescale of adjustment depends on mantle viscosity and the spatial scale of the disturbance. Post-glacial rebound operates over thousands to tens of thousands of years, while responses to large-scale tectonic loading can take millions of years.

Models of Isostasy

Airy and Pratt Isostatic Models

The Airy model assumes the crust has a uniform density but varies in thickness. Topographic highs are compensated by deep crustal roots extending into the mantle, while topographic lows (like ocean basins) have thinner crust. Think of wooden blocks of different heights floating in water: taller blocks stick up higher but also extend deeper below the waterline.

- The compensation condition can be written as: $\rho_c \cdot t = \rho_m \cdot r$, where ρ_c is crustal density, t is the elevation above a reference, ρ_m is mantle density, and r is the root depth below the reference Moho.
- This model works well in regions with large crustal roots, such as the Himalayas and the Andes, where seismic data confirm thickened crust beneath high topography.

The Pratt model assumes the crust has a uniform thickness but varies in density. Mountains are underlain by less dense crust, while lowlands and ocean floors are underlain by denser crust. The compensation depth (base of all columns) is the same everywhere.

- Mathematically: $\rho_1 \cdot h_1 = \rho_2 \cdot h_2 = \text{constant}$ for all columns extending to the compensation depth.
- This model applies well in regions with lateral density variations, such as the Basin and Range Province (where thermal expansion reduces crustal density beneath elevated terrain) and the East African Rift System.

Flexural Isostatic Model and Combined Approaches

The Vening Meinesz (flexural) model treats the lithosphere as an elastic plate that bends under applied loads rather than responding purely locally. A point load doesn't just push the crust down directly beneath it; the rigidity of the plate spreads the deflection over a wider area, creating a depression under the load flanked by a subtle peripheral bulge.

- The key parameter is flexural rigidity (D), which depends on the elastic thickness (T_e) of the lithosphere: $D = \frac{ET_e^3}{12(1-\nu^2)}$, where E is Young's modulus and ν is Poisson's ratio.
- This model explains features like the moat and arch around the Hawaiian Islands (where the volcanic load bends the Pacific plate) and the broad rebound pattern of the Fennoscandian Shield.

In practice, Earth's isostatic behavior combines elements of all three models. The relative importance of each depends on the geological setting and spatial scale. At very large scales (hundreds of km), local Airy or Pratt compensation dominates. At intermediate scales, flexural effects become significant. Combined approaches like the Airy-Heiskanen and Pratt-Hayford models incorporate elements of both end-member models for more realistic representations. Choosing the right model depends on the tectonic setting, crustal age and composition, and available geophysical constraints.

Isostasy and Gravity Anomalies

Gravity Anomalies and Their Relationship to Isostasy

A gravity anomaly is the difference between observed gravity at a point and the theoretical gravity predicted by a reference model (typically the reference ellipsoid). Because isostatic compensation redistributes mass at depth, it directly shapes the gravity field measured at the surface.

- A positive gravity anomaly indicates excess mass beneath the observation point, which could mean dense rocks near the surface or topography that isn't fully compensated by a crustal root.
- A negative gravity anomaly indicates a mass deficit, such as a thick low-density crustal root or overcompensated topography.

If a mountain range is perfectly isostatically compensated, the extra mass of the topography above is offset by the low-density root below, and the gravity anomaly (after appropriate corrections) should be near zero. Departures from zero tell you the region is out of isostatic equilibrium.

Types of Gravity Anomalies and Their Calculation

Each type of gravity anomaly removes different effects, progressively isolating subsurface structure:

1.

Free-air anomaly: Start with observed gravity, then apply the free-air correction to account for the station's elevation above the reference ellipsoid. This anomaly reflects the total mass distribution beneath the observation point but doesn't remove the gravitational effect of topographic mass itself.

2.

Bouguer anomaly: Take the free-air anomaly and apply the Bouguer correction, which removes the gravitational attraction of the rock mass between the station and the reference level (plus terrain corrections for irregular topography). The result highlights subsurface density variations. Over continents, Bouguer anomalies are typically negative because the correction removes topographic mass but the compensating root remains as a mass deficit.

3.

Isostatic anomaly: Take the Bouguer anomaly and remove the calculated gravitational effect of the assumed isostatic compensation (using an Airy, Pratt, or flexural model). What's left reveals departures from isostatic equilibrium. A near-zero isostatic anomaly means the chosen model fits well.

Notable examples: the Tibetan Plateau shows a strongly positive free-air anomaly (high elevation with large total mass) but a strongly negative Bouguer anomaly (thick, low-density crustal root). The Mid-Atlantic Ridge shows a negative Bouguer anomaly due to hot, low-density mantle material beneath the ridge.

Applying Isostatic Principles

Crustal Thickness and Density Estimation

By assuming an isostatic model and using observed topography and gravity data, you can estimate crustal thickness and density variations across a region. For example, under the Airy model, you can invert topographic data to predict Moho depth. These gravity-derived estimates are then validated against independent seismic data (from seismic refraction surveys or receiver function analysis). Where the two agree, confidence in the crustal model is high. Where they disagree, it points to regions that may be out of isostatic equilibrium or where the chosen model is too simple.

Vertical Motions and Sedimentary Basin Evolution

Isostatic models predict how the crust responds vertically to changes in surface loading:

- Post-glacial rebound: After ice sheets melt, the unloaded crust rises. GPS measurements in Scandinavia, Canada, and Antarctica track this ongoing uplift, and the rate of rebound provides constraints on mantle viscosity (typically 10^{20} to 10^{21} Pa·s for the upper mantle).

- Sedimentary isostasy: As sediments accumulate in a basin, their weight causes further subsidence, which creates more accommodation space for additional sediment. This positive feedback loop explains why basins like the Gulf Coast and North Sea can accumulate sedimentary sequences many kilometers thick. The relationship is roughly that for every 1 km of sediment added, the basin floor subsides by an additional amount depending on the density contrast: $w \approx h \cdot \frac{\rho_s}{\rho_m - \rho_s}$, where ρ_s is sediment density and ρ_m is mantle density.

Mountain Building and Geodynamic Modeling

During orogenesis (mountain building), crustal thickening from collision or compression triggers isostatic uplift. The Tibetan Plateau (~5 km average elevation) is supported by a crustal root extending to ~70 km depth, roughly double the global continental average of ~35 km. The Altiplano-Puna Plateau in the Central Andes shows a similar pattern.

Isostatic compensation is built into numerical geodynamic models that simulate long-term lithospheric deformation. Thin-sheet models and finite element models incorporate isostatic response to tectonic loading, allowing researchers to test hypotheses about the forces driving crustal evolution. Without isostasy, these models would predict unrealistic topography and gravity fields.

3.4 Geodetic techniques and their applications

Geodetic Techniques and Principles

Geodetic Techniques and Measurements

Geodetic techniques are the tools geophysicists use to precisely measure Earth's shape, size, orientation, and gravity field. Each technique captures different aspects of the planet's geometry and dynamics, and understanding what each one does well (and where it falls short) is central to interpreting geophysical data.

Leveling measures height differences between points on Earth's surface using a surveyor's level and graduated rods. It's the classic method for establishing vertical control networks and determining orthometric heights (heights referenced to the geoid, or mean sea level surface). Leveling is highly accurate over short distances but becomes time-consuming and error-prone over long baselines.

Triangulation determines horizontal positions by measuring angles between visible points. You need a known baseline distance and at least one astronomically determined azimuth to anchor the network. This was the backbone of national survey networks for centuries before satellite methods took over.

Trilateration works differently: instead of angles, it measures distances between points using electronic distance measurement (EDM) instruments. When combined with angular measurements, trilateration establishes horizontal control networks and determines point coordinates. EDM made trilateration far more practical than traditional baseline measurements.

Principles and Technologies Behind Geodetic Techniques

Global Positioning System (GPS) uses a constellation of satellites to determine precise 3D positions on Earth's surface. The core principle is trilateration in three dimensions: a GPS receiver measures the travel time of signals from multiple satellites, then converts those times to distances using the speed of light. Atomic clocks on the satellites ensure timing accuracy at the nanosecond level, which matters because a 1-nanosecond error translates to roughly 30 cm of position error.

Interferometric Synthetic Aperture Radar (InSAR) uses satellite-based radar to measure surface deformation with millimeter-level accuracy. Here's how it works:

1. A radar satellite images the same area at two different times.
2. Each pixel in the radar image records both amplitude and phase of the returned signal.
3. The phase difference between the two images reveals how much the ground surface moved toward or away from the satellite between acquisitions.
4. These phase differences are mapped across the scene to produce an interferogram, which shows the spatial pattern of deformation.

Gravimetry measures Earth's gravity field using ground-based gravimeters or satellite missions like GRACE (Gravity Recovery and Climate Experiment). Gravity measurements serve multiple purposes: determining the shape of the geoid, detecting subsurface density variations, and tracking mass redistribution over time.

Applications of Geodetic Techniques in Geophysics

Monitoring Earth's Surface Deformation and Tectonic Processes

Geodetic techniques provide the primary observational constraints on how Earth's surface moves and deforms, which feeds directly into models of tectonic and volcanic processes.

- Leveling and GPS detect vertical land movements tied to tectonic activity, including coseismic uplift or subsidence during earthquakes, interseismic strain accumulation, and inflation/deflation of volcanic edifices.
- InSAR maps surface deformation caused by earthquakes, volcanic eruptions, landslides, and ground subsidence over broad areas. For example, InSAR interferograms of the 1999 Izmit earthquake in Turkey revealed the detailed slip distribution along the North Anatolian Fault. This kind of spatial coverage is something point-based methods like GPS simply cannot match.
- GPS and InSAR together monitor slow deformation processes such as tectonic plate motion (typically a few cm/year), postglacial rebound, and fault creep. These long-term velocity measurements provide critical constraints for geodynamic models.

Studying Earth's Internal Structure and Resource Exploration

- Ground-based gravity measurements detect subsurface density contrasts associated with geological structures like sedimentary basins (low density), igneous intrusions (high density), and ore deposits. Gravity surveys remain a standard tool in mineral and hydrocarbon exploration, as well as crustal structure studies.

- Satellite gravimetry, particularly the GRACE and GRACE-FO missions, tracks temporal changes in Earth's gravity field caused by mass redistribution. GRACE data have revealed rates of ice sheet mass loss in Greenland and Antarctica (on the order of hundreds of gigatons per year), mapped groundwater depletion in regions like northern India and California's Central Valley, and measured ongoing postglacial rebound in Scandinavia and Canada. These observations have direct implications for sea-level rise projections and water resource management.

Interpretation of Geodetic Data

Understanding Techniques, Limitations, and Potential Errors

Every geodetic dataset carries assumptions and error sources that you need to understand before interpreting results.

- Leveling data give height differences between benchmarks. They're used to establish vertical datums, study land subsidence or uplift, and support engineering projects. The main limitations are slow acquisition speed and susceptibility to systematic errors that accumulate over long lines (rod calibration errors, refraction effects).

- Triangulation and trilateration data provide horizontal coordinates of control points. These are used to build geodetic reference frames, create topographic maps, and study crustal deformation. Accuracy depends on line-of-sight visibility and atmospheric conditions affecting EDM measurements.

- GPS data provide precise 3D positions and velocities. Modern continuous GPS networks achieve millimeter-level accuracy in position and sub-mm/year accuracy in velocity. Sources of error include tropospheric and ionospheric signal delays, multipath reflections, and satellite orbit uncertainties.

Analyzing and Interpreting Geodetic Datasets

- InSAR data produce spatially continuous maps of surface deformation, but interpretation requires care. Atmospheric phase delays (especially water vapor variations) can mimic or obscure real deformation signals. Temporal decorrelation in vegetated areas reduces data quality. Despite these challenges, InSAR interferograms can be modeled to estimate fault parameters (geometry, slip magnitude, rake) or magma chamber properties (depth, volume change).

- Gravity data reveal information about Earth's gravity field and subsurface density structure. After applying standard corrections (free-air, Bouguer, terrain), gravity anomalies can be used to model the geoid, estimate crustal thickness, and identify density anomalies tied to geological structures. The inherent non-uniqueness of gravity inversion means that multiple subsurface models can fit the same gravity data, so independent constraints from other methods are often essential.

Integration of Geodetic and Geophysical Data

Combining Geodetic Data with Other Geophysical Datasets

No single technique tells the whole story. Integrating geodetic data with other geophysical observations produces a far more complete picture of Earth's structure and dynamics.

- Deformation + seismology for earthquake studies: InSAR and GPS data constrain fault geometry and the spatial distribution of slip, while seismic data reveal the rupture process, energy release, and temporal evolution. Together, they allow geophysicists to build kinematic and dynamic models of earthquake sources.

- Gravity + seismic + magnetic + geological data for crustal modeling: Gravity anomalies constrain density structure, seismic velocities constrain elastic properties, and magnetic data reveal variations in rock magnetization. Combining these datasets in joint inversions produces 3D crustal models that are better constrained than any single-method model. This approach is widely used to identify sedimentary basins, igneous intrusions, and mineral deposits.

Advancing Earth Science Through Data Integration

- Plate motion and mantle dynamics: GPS and InSAR velocity fields, combined with geodynamic models, help constrain the driving forces of plate tectonics and mantle convection. These integrated models provide estimates of mantle viscosity and lithospheric rheology that cannot be obtained from surface observations alone.

- Mass transport and climate: Satellite gravimetry data (GRACE/GRACE-FO) are integrated with satellite altimetry, ice-penetrating radar, and ground-based hydrological measurements to study Earth's mass transport processes. This includes ice sheet dynamics, sea-level change, and hydrological cycles. The integration reveals how the solid Earth, oceans, ice sheets, and atmosphere interact, providing quantitative constraints on the impacts of climate change.

Unit 4 – Geomagnetism

4.1 Earth's magnetic field and its origin

Earth's Magnetic Field: Characteristics and Components

Earth's magnetic field originates deep within the planet's liquid outer core, where convective motion of electrically conducting iron generates a self-sustaining dynamo. This field shields the surface from harmful solar radiation, provides the basis for magnetic navigation, and preserves a record of tectonic and geologic history in magnetized rocks. The sections below cover the field's structure, its generation mechanism, how it changes over time, and how it interacts with the solar wind.

Dipole Field and Magnetic Poles

To a first approximation, Earth's magnetic field resembles that of a bar magnet (a dipole) tilted about 11° from the rotational axis. Field lines emerge from the south magnetic pole and converge at the north magnetic pole. Two angles describe the field's orientation at any point on the surface:

- Declination is the angle between geographic north and magnetic north. It varies with location and changes slowly over time.
- Inclination (or dip) is the angle the field vector makes with the horizontal. At the magnetic equator, inclination is 0° (field lines run parallel to the surface). At the magnetic poles, inclination is ±90° (field lines point straight into or out of the ground).

Field Strength and Composition

Surface field strength ranges from about 25,000 nT near the magnetic equator to roughly 65,000 nT near the poles. The total observed field is a superposition of three contributions:

- Main field (~95% of surface strength): Generated by dynamo action in the liquid outer core.
- Crustal (or lithospheric) field: Produced by permanently magnetized minerals in the crust. This component is relatively weak but carries important information about past field directions.
- External field: Arises from electric current systems in the ionosphere and magnetosphere, driven by solar wind interactions. It fluctuates on timescales of seconds to days.

Dynamo Theory: Generating Earth's Magnetic Field

Convection Currents and Electrical Conductivity

The geodynamo theory explains how Earth maintains a planetary-scale magnetic field. The liquid outer core is composed primarily of iron and nickel, making it an excellent electrical conductor. Vigorous convection in this layer is driven by two main energy sources:

1. Thermal convection from heat released at the inner-outer core boundary as the solid inner core slowly crystallizes.
2. Compositional convection from the release of light elements (such as sulfur, oxygen, and silicon) during inner-core solidification. These buoyant fluids rise through the denser surrounding liquid, sustaining flow.

Self-Sustaining Geodynamo

The dynamo becomes self-sustaining through a feedback loop:

1. Convective motions move electrically conducting fluid through an existing (even weak) magnetic field.

2. By electromagnetic induction, these motions generate electric currents in the fluid.

3. Those electric currents, in turn, produce magnetic fields that reinforce and reshape the original field.

4. Earth's rotation imposes a Coriolis force on the convecting fluid, organizing the flow into roughly columnar patterns aligned with the spin axis. This is why the resulting field is predominantly dipolar and aligned close to the rotation axis.

The geodynamo will persist as long as there is enough heat flux from the inner core and a continuing supply of buoyant light elements to drive outer-core convection.

Temporal Variations of Earth's Magnetic Field

Secular Variation

Secular variation refers to gradual changes in the field's strength, direction, and pole positions over timescales of years to centuries. These changes reflect evolving flow patterns in the liquid outer core. Observable examples include:

- The north magnetic pole has been migrating from the Canadian Arctic toward Siberia, accelerating in recent decades to roughly 50 km/yr.

- Global dipole moment has decreased by about 10% over the past 150 years, though this does not necessarily signal an imminent reversal.

Geomagnetic Reversals

At irregular intervals, the dipole field reverses polarity entirely: the north and south magnetic poles swap. Key facts about reversals:

- The time between successive reversals varies enormously, from tens of thousands to tens of millions of years. The most recent full reversal, the Brunhes-Matuyama reversal, occurred approximately 780,000 years ago.

- A reversal is not instantaneous. The transition typically takes several thousand years, during which field intensity drops significantly and the field geometry becomes complex and non-dipolar.

- The reversal record is preserved in rocks. As new oceanic crust forms at mid-ocean ridges, iron-bearing minerals lock in the ambient field direction as they cool through their Curie temperature. This produces symmetric magnetic stripes on either side of the ridge, which were a key piece of evidence for seafloor spreading and plate tectonics. On land, stacked lava flows record the same polarity history, forming the basis of magnetic stratigraphy used to date sedimentary and volcanic sequences.

Earth's Magnetic Field and the Solar Wind

Magnetosphere and Solar Wind Interaction

The solar wind is a continuous stream of charged particles (mostly electrons and protons) flowing outward from the Sun's corona at speeds of 300–800 km/s. Earth's magnetic field deflects most of these particles, carving out a cavity in the solar wind called the magnetosphere.

- On the dayside (Sun-facing), solar wind pressure compresses the magnetosphere; the boundary (the magnetopause) typically sits about 10 Earth radii from the surface.
- On the nightside, the field is stretched into a long magnetotail extending hundreds of Earth radii downstream.

Geomagnetic Storms and Auroras

During intense solar events such as coronal mass ejections (CMEs), a surge of plasma and enhanced magnetic fields strikes the magnetosphere. This can trigger geomagnetic storms, which pose real hazards:

- Induced currents can overload power grids (the 1989 Quebec blackout is a well-known example).
- Satellite electronics, GPS accuracy, and high-frequency radio communications can all be degraded.

Auroras (the northern and southern lights) are a visible consequence of solar wind-magnetosphere coupling. Solar wind particles that enter the magnetosphere are funneled along field lines toward the polar regions. When these energetic particles collide with atmospheric gases, the gases emit light: oxygen produces green and red hues, while nitrogen contributes blue and purple. Auroras are typically confined to oval-shaped zones around the magnetic poles, but during strong geomagnetic storms they can be visible at much lower latitudes.

4.2 Magnetic properties of rocks and minerals

Magnetic properties of rocks and minerals are central to understanding Earth's magnetic field, its history, and the dynamics of the planet's interior. These properties originate from the behavior of electrons in atoms, which produce magnetic responses ranging from negligibly weak to strong enough to preserve a record of ancient field directions.

Iron-bearing minerals, especially magnetite, are the main carriers of magnetism in rocks. Because these minerals can retain magnetization over geologic time, they serve as archives of Earth's past magnetic field. This makes them indispensable for studying geomagnetic reversals, tectonic plate movements, and core dynamics.

Magnetic Material Types

Diamagnetic, Paramagnetic, and Ferromagnetic Materials

All materials respond to an applied magnetic field, but the strength and character of that response varies enormously depending on electronic structure.

- Diamagnetic materials have a weak, negative magnetic susceptibility and are slightly repelled by a magnetic field. All electron magnetic moments cancel out, so there are no unpaired electrons contributing a net moment. They do not retain any magnetization once the field is removed. Common examples: quartz, calcite, water.

- Paramagnetic materials have a small, positive magnetic susceptibility and are weakly attracted to a magnetic field. They contain unpaired electrons whose moments align parallel to an applied field, but the alignment is easily disrupted by thermal energy. They also lose their magnetization when the field is removed. Common examples: biotite, pyrite, siderite.
- Ferromagnetic (sensu lato) materials have a large, positive magnetic susceptibility and are strongly attracted to a magnetic field. Neighboring magnetic moments interact strongly and align spontaneously within regions called magnetic domains. Crucially, these materials can retain magnetization after the external field is removed. This retained magnetization is called remanent magnetization. Common examples: magnetite, hematite, pyrrhotite.

A note on terminology: in geophysics, "ferromagnetic" is often used loosely to include ferrimagnetic materials (like magnetite, where opposing sublattice moments don't fully cancel) and antiferromagnetic materials with a parasitic moment (like hematite). True ferromagnetism, where all moments align in parallel, is seen in pure iron but is rare in natural minerals.

Factors Determining Magnetic Behavior

The magnetic behavior of a material comes down to the presence and alignment of magnetic moments from unpaired electrons:

- In diamagnetic materials, electrons are all paired, so their orbital and spin moments cancel. The only response is a tiny induced moment opposing the applied field.
- In paramagnetic materials, unpaired electrons create net atomic moments, but thermal agitation keeps them randomly oriented unless an external field is applied. The resulting alignment is weak and temporary.
- In ferromagnetic materials, strong exchange interactions between neighboring atoms cause spontaneous alignment of moments within domains. This cooperative behavior produces a much larger magnetization that can persist without an external field.

Magnetic Properties of Materials

Magnetic Susceptibility and Remanent Magnetization

Magnetic susceptibility (χ) measures how readily a material becomes magnetized in an external field. It is defined as the ratio of induced magnetization (M) to the applied field strength (H):

$$\chi = \frac{M}{H}$$

In SI units, χ is dimensionless (though volume and mass susceptibility have different unit conventions). Susceptibility values span many orders of magnitude: diamagnetic minerals like quartz have χ on the order of -10^{-5}, paramagnetic minerals like biotite around 10^{-4} to 10^{-3}, and magnetite can reach values above 1.

Remanent magnetization (or remanence) is the permanent magnetization that remains in a material after the external field is removed. Only ferromagnetic (sensu lato) minerals carry remanence, because their domain structure can lock in a preferred alignment of moments. This remanence is what makes paleomagnetism possible.

Magnetic Anisotropy

Magnetic anisotropy refers to the directional dependence of a material's magnetic properties. A rock's susceptibility may differ depending on the direction in which the field is applied. Several factors cause this:

- Shape anisotropy: elongated magnetic grains are easier to magnetize along their long axis.
- Crystalline anisotropy: the crystal lattice of a mineral has preferred "easy" magnetization directions.
- Stress/strain anisotropy: tectonic deformation can reorient grains or alter domain structures.

The anisotropy of magnetic susceptibility (AMS) is a widely used technique that measures susceptibility in multiple directions to define an ellipsoid. This ellipsoid reveals the preferred orientation and alignment of magnetic minerals, providing information about rock fabric and deformation history. For example, AMS can identify sedimentary bedding orientations, metamorphic foliation, and flow directions in igneous rocks.

Magnetic Properties of Rocks

Contribution of Rock-Forming Minerals

The magnetic character of a rock is dominated by its iron-bearing minerals:

- Magnetite (Fe_3O_4): The most important magnetic mineral in geophysics. It is ferrimagnetic with a very high susceptibility and can carry extremely stable remanent magnetization. Even small concentrations of magnetite dominate a rock's magnetic signature.
- Hematite (α-Fe_2O_3): Antiferromagnetic with a weak parasitic ferromagnetic moment due to spin canting. Its susceptibility is much lower than magnetite's, but it can carry very stable remanence over billions of years, making it valuable for paleomagnetic studies of old rocks.
- Pyrrhotite ($Fe_{1-x}S$): Ferrimagnetic with strong susceptibility, but its magnetic properties depend on composition (the value of x) and crystal structure. It is an important carrier in some sulfide-rich rocks.
- Goethite (α-FeOOH): A common weathering product with weak magnetic properties, sometimes relevant in laterites and soils.

Factors Influencing Rock Magnetic Properties

Several factors control a rock's overall magnetic behavior:

- Mineral concentration: More ferromagnetic mineral content means a stronger magnetic signature. Mafic igneous rocks like basalts and gabbros are typically strongly magnetic because they are rich in magnetite.
- Grain size: This profoundly affects magnetic stability. Very fine grains (below ~0.1 μm for magnetite) are single-domain (SD), meaning the entire grain is one magnetic domain. SD grains carry the most stable remanence. Larger grains are multi-domain (MD) and have lower coercivity and less stable remanence. An intermediate range called pseudo-single-domain (PSD) shows properties between the two.
- Mineral distribution: Whether magnetic grains are evenly dispersed or clustered affects the bulk magnetic response and how grains interact magnetically.
- Secondary processes: Alteration, weathering, and metamorphism can create new magnetic minerals, destroy existing ones, or change grain sizes. These processes modify both susceptibility and remanence, sometimes overprinting the original magnetic record.

Rocks dominated by diamagnetic or paramagnetic minerals, such as limestones and quartzites, have very weak magnetic signatures.

Magnetic Remanence in Rocks

Acquisition Processes

Rocks acquire remanent magnetization through several distinct mechanisms, each tied to a specific geological process:

1.

Thermoremanent magnetization (TRM): Acquired when ferromagnetic minerals cool through their Curie temperature (T_C) in the presence of an ambient magnetic field. Above T_C, thermal energy overcomes exchange interactions and the material is paramagnetic. As the rock cools below T_C, magnetic moments align with the field and become locked in. TRM is the primary remanence in igneous rocks and high-grade metamorphic rocks. For magnetite, $T_C \approx 580°C$; for hematite, $T_C \approx 675°C$.

2.

Chemical remanent magnetization (CRM): Acquired when new ferromagnetic minerals nucleate and grow through a critical grain size in the presence of a magnetic field. As a grain grows past the superparamagnetic threshold, its moment becomes locked in the ambient field direction. CRM forms during diagenesis, hydrothermal alteration, or oxidation (e.g., magnetite altering to hematite).

3.

Detrital remanent magnetization (DRM): Acquired when magnetic mineral grains physically rotate to align with Earth's field during deposition in water. As sediment settles and compacts, the alignment is preserved. This is the primary remanence mechanism in clastic sedimentary rocks like sandstones and mudstones.

4.

Post-depositional remanent magnetization (pDRM): Acquired after deposition but before full lithification. In water-saturated, unconsolidated sediment, magnetic grains can still rotate within pore spaces and align with the ambient field. pDRM often provides a more accurate record of the field at the time of early burial than DRM, because it smooths out mechanical misalignment during settling.

Preservation and Significance

The usefulness of remanent magnetization depends on how well it is preserved. Preservation requires:

- Stable magnetic minerals with high coercivity (resistance to demagnetization), such as single-domain magnetite or fine-grained hematite.
- Minimal thermal overprinting: If a rock is reheated above the Curie temperature of its magnetic minerals, the original remanence is reset.
- Minimal chemical alteration: Growth of new magnetic phases can overprint or obscure the primary signal.
- No exposure to strong external fields: Lightning strikes, for example, can locally remagnetize surface rocks.

Rocks that retain their original remanence are the foundation of paleomagnetism. By measuring the direction and intensity of remanence in dated rock samples, researchers can:

- Reconstruct apparent polar wander paths, which track the movement of tectonic plates relative to the magnetic pole over time.

- Build the geomagnetic polarity timescale, documenting the history of field reversals recorded in ocean-floor basalts and sedimentary sequences.
- Constrain paleogeographic reconstructions, determining where continents were located in the geological past.
- Study secular variation and the long-term evolution of Earth's core dynamics.

4.3 Magnetic surveying and data interpretation

Magnetic Surveying Principles and Techniques

Principles and Instrumentation

Magnetic surveying measures variations in Earth's magnetic field to detect subsurface geological features and mineral deposits. Because the magnetic field is a vector quantity with both magnitude and direction, these variations carry information about what lies beneath the surface.

Ground magnetic surveys use portable magnetometers carried along traverses or across a grid pattern. The two most common instrument types are:

- Proton precession magnetometers measure total field intensity by detecting the precession frequency of protons in a hydrogen-rich fluid. They're robust and require no orientation but sample discretely rather than continuously.
- Fluxgate magnetometers measure individual components of the field vector. They can sample continuously and are sensitive to field direction, but they need careful orientation.

Airborne magnetic surveys mount magnetometers on fixed-wing aircraft or helicopters, allowing rapid coverage of large areas. These surveys are typically flown along parallel lines at a constant elevation above terrain.

Survey Design and Data Collection

Magnetic data are recorded as total magnetic intensity (TMI) or as individual components of the field vector (horizontal, vertical, or total). Several design parameters control the quality and cost of a survey:

- Line spacing and sampling interval determine spatial resolution. Closer spacing gives finer detail but increases time and cost.
- Survey altitude controls sensitivity to shallow sources. Lower altitudes enhance detection of near-surface features, while higher altitudes smooth out shallow noise and emphasize deeper, broader anomalies.

The right balance depends on target depth and the scale of features you're trying to resolve.

Magnetic Anomalies and Subsurface Geology

Magnetic Properties of Rocks and Minerals

Magnetic anomalies are local deviations from the expected background field, caused by magnetic minerals in the subsurface. The most important of these minerals is magnetite (Fe_3O_4). Anomalies can be:

- Positive: magnetic intensity higher than the background field
- Negative: magnetic intensity lower than the background field

The key physical property controlling anomaly strength is magnetic susceptibility, which describes how easily a material becomes magnetized in an external field. Ferromagnetic minerals like magnetite have very high susceptibility and are the primary sources of magnetic anomalies.

As a general rule, igneous and metamorphic rocks tend to have higher magnetic susceptibility than sedimentary rocks, making them more likely to generate detectable anomalies. This is because crystallization and metamorphic processes concentrate magnetite and other iron-oxide minerals.

Geologic Structures and Mineral Deposits

The shape, amplitude, and wavelength of a magnetic anomaly encode information about the geometry, depth, and magnetic properties of the source body.

- Mineral deposits containing magnetic minerals (e.g., iron ore with abundant magnetite) produce distinct anomalies that aid detection and delineation.
- Faults juxtapose rocks with different magnetic properties and often appear as linear features or sharp offsets in the magnetic data.
- Folds can produce characteristic "bulls-eye" patterns (for domes or basins) or arcuate anomalies that trace the fold axis.
- Intrusions of mafic or ultramafic composition typically stand out as strong positive anomalies against less magnetic country rock.

Data Processing for Magnetic Surveys

Raw magnetic data contain signals from both the geology you care about and several sources of noise. Processing strips away the noise and reshapes the data to make interpretation more straightforward.

Diurnal Correction

Earth's external magnetic field fluctuates throughout the day due to ionospheric currents driven by solar activity. These diurnal variations can reach tens of nanoteslas and must be removed.

1. Set up a base station magnetometer at a fixed location within or near the survey area.
2. Record the field continuously at the base station throughout the survey day.
3. Subtract the time-varying base station signal from each survey reading, matched by timestamp.

This isolates the spatial variations caused by geology from the temporal variations caused by solar activity.

Reduction-to-Pole (RTP)

At most latitudes, the Earth's magnetic field is inclined rather than vertical. This inclination causes magnetic anomalies to be asymmetric and shifted away from directly above their source bodies. Reduction-to-pole is a mathematical transformation that recalculates what the anomaly pattern would look like if the field were perfectly vertical (as at the magnetic pole).

After RTP, anomalies are centered over their causative bodies, which makes interpretation much more intuitive. RTP works well at mid to high latitudes but becomes unstable near the magnetic equator where inclination approaches zero.

Additional Processing Techniques

- Leveling removes systematic differences between adjacent flight lines caused by heading errors or drift.
- Gridding interpolates irregularly spaced data onto a regular grid for map display.
- Filtering enhances or suppresses features at different spatial scales:
- Low-pass filters smooth the data, emphasizing broad, deep sources.
- High-pass filters sharpen the data, highlighting shallow or small-scale features.
- Band-pass filters isolate anomalies within a specific wavelength range.

Magnetic data are often integrated with other geophysical datasets (gravity, electromagnetic) and geological information such as drill hole logs, geologic maps, and seismic sections. This integration constrains the interpretation and reduces the inherent ambiguity of potential field methods.

Interpreting Magnetic Data for Geology and Minerals

Magnetic Anomaly Maps and Profiles

Anomaly maps display the spatial distribution of magnetic field variations across the survey area using color scales or contour lines. They let you identify regional trends, locate discrete anomalies, and recognize geological patterns like fault networks or lithological contacts.

Magnetic profiles plot field intensity along a single survey line, giving a cross-sectional view. Profiles are especially useful for analyzing the shape, amplitude, and wavelength of individual anomalies, which feed directly into depth and geometry estimates.

Interpretation Techniques and Considerations

Interpreting anomalies involves linking their characteristics to plausible geological sources:

- Positive anomalies often indicate highly magnetic lithologies such as mafic intrusions (e.g., gabbro) or iron-rich ore bodies.
- Negative anomalies may indicate less magnetic rocks (e.g., sedimentary units) or hydrothermal alteration zones where magnetite has been destroyed.

Anomaly shape carries geometric information:

- Symmetric, circular anomalies suggest vertical or steeply dipping bodies, such as kimberlite pipes or vertical plugs.
- Elongated or asymmetric anomalies point to dipping structures like inclined dikes or fault-bounded blocks.

Depth estimation techniques relate anomaly shape to source depth:

- The half-width rule uses the horizontal distance from the anomaly peak to the point where intensity drops to half its maximum value. For a simple sphere, the depth to center is approximately $z \approx 1.3 \times x_{1/2}$, where $x_{1/2}$ is the half-width.
- Euler deconvolution is an automated method that solves for source location and depth using the spatial derivatives of the field, with a "structural index" chosen to match the expected source geometry

(e.g., 0 for a contact, 1 for a dike, 2 for a horizontal cylinder, 3 for a sphere).

As with all potential field methods, magnetic interpretation suffers from non-uniqueness: multiple subsurface configurations can produce the same surface anomaly. Integrating magnetic results with gravity data, geological mapping, and drill hole information is the best way to narrow down the most geologically reasonable model.

4.4 Paleomagnetism and its applications

Paleomagnetism and Rock Magnetism

Principles of Paleomagnetism

Paleomagnetism studies Earth's ancient magnetic field as preserved in rocks. When rocks form or cool below their Curie temperature, magnetic minerals lock in a record of the field's direction and intensity at that moment. This "fossil magnetism" gives us a direct window into what Earth's magnetic field looked like millions or even billions of years ago.

The permanent magnetization rocks acquire this way is called natural remanent magnetization (NRM), and it comes in several flavors depending on how the rock formed:

- Thermal remanent magnetization (TRM): Acquired when magnetic minerals in igneous rocks cool below their Curie temperature in the ambient field. This also applies to sedimentary rocks baked by nearby igneous intrusions. TRM tends to be the most stable and reliable form of NRM.

- Chemical remanent magnetization (CRM): Acquired when new magnetic minerals form or grow through chemical processes (diagenesis, metamorphism) in the presence of an ambient field. Because the minerals crystallize at lower temperatures, CRM can sometimes record a field direction different from the original formation age.

- Detrital remanent magnetization (DRM): Acquired when magnetic mineral grains physically align with Earth's field as they settle through water and become locked into sedimentary layers. DRM is common in fine-grained sediments like red beds and deep-sea clays.

Natural Remanent Magnetization (NRM)

NRM is the foundation of all paleomagnetic work. Its usefulness depends entirely on how faithfully it preserves the original field signal.

The stability of NRM is controlled by several factors:

- Magnetic mineral composition: Magnetite (Fe_3O_4) and hematite (α-Fe_2O_3) are the most common carriers. Magnetite has a Curie temperature of ~580°C and tends to carry strong, stable remanence. Hematite has a higher Curie temperature (~675°C) but weaker magnetization.

- Grain size: Single-domain and pseudo-single-domain grains hold the most stable remanence. Large multi-domain grains are more susceptible to remagnetization.

- Secondary overprints: Later thermal or chemical events can partially or completely reset the original NRM. Paleomagnetic lab techniques (progressive thermal or alternating-field demagnetization) are used to strip away these overprints and isolate the primary magnetization.

Rocks with particularly stable NRM include basalts, volcanic tuffs, and well-preserved sedimentary rocks such as red beds and certain limestones.

Apparent Polar Wander and Plate Tectonics

Apparent Polar Wander (APW)

If you measure paleomagnetic directions from rocks of different ages on a single continent, the calculated pole positions appear to move through time. This phenomenon is called apparent polar wander (APW).

The key insight: the magnetic poles haven't actually migrated across the globe. Instead, the continent itself has moved relative to Earth's spin axis. The "wandering" is apparent because we're viewing the pole from a reference frame fixed to a moving plate.

APW paths are constructed by plotting paleopole positions (ancient magnetic pole locations) from progressively older rocks on a single continent. Each continent produces its own APW path. For example, the North American APW path and the European APW path diverge going back in time, reflecting the opening of the Atlantic Ocean.

Plate Tectonics and Paleomagnetism

APW paths provided some of the most compelling early evidence for plate tectonics:

- When APW paths from two continents that were once joined (say, North America and Europe before the Atlantic opened) are reconstructed, they converge into a single path for the period when those continents were together. This convergence is strong evidence for past supercontinents like Pangaea and Rodinia.
- Differences between APW paths from separate continents reveal when and how those continents split apart and drifted to their current positions.
- Paleomagnetic data have been instrumental in reconstructing the breakup of Pangaea (~200 Ma), tracing the separation of Africa and South America, the northward drift of India, and the assembly of earlier supercontinents.

Paleomagnetic Data for Reconstruction and Geodynamo Studies

Continental Reconstruction

Paleomagnetic measurements give two pieces of information that are directly useful for placing continents on the ancient globe:

1. Inclination tells you the paleolatitude. Earth's magnetic field geometry means that inclination varies systematically with latitude. The relationship is given by the dipole formula: $\tan(I) = 2\tan(\lambda)$, where I is the magnetic inclination and λ is the latitude. Steep inclinations indicate high latitudes; shallow inclinations indicate positions near the equator.
2. Declination tells you the rotation of the continent relative to magnetic north. If a continent has rotated clockwise since the rock formed, the recorded declination will be offset from the expected north direction.

By comparing paleolatitudes and declinations from rocks of the same age on different continents, you can reconstruct their relative positions. This approach has been used to map out the configurations of Gondwana, Laurasia, and the full Pangaea supercontinent.

One important limitation: paleomagnetic data constrain paleolatitude and rotation but not paleolongitude. Determining east-west positions requires additional geological or hotspot-track evidence.

Geodynamo Studies

Beyond plate reconstructions, paleomagnetic records also reveal how Earth's magnetic field generator (the geodynamo) has behaved over deep time.

- Field reversals: Paleomagnetic data show that Earth's magnetic field periodically flips polarity, with magnetic north and south switching places. These reversals are not periodic; the intervals between them range from tens of thousands to tens of millions of years. The most recent major reversal is the Brunhes-Matuyama reversal (~780 ka), and shorter events like the Jaramillo subchron (~1.07–1.00 Ma) are also well documented.
- Paleointensity: The strength of the ancient field can be estimated from the intensity of NRM (using techniques like the Thellier method for TRM-bearing rocks). These records show that field intensity drops significantly during reversals and varies by factors of 2–5 over millions of years.
- Secular variation: Shorter-term fluctuations in field direction and intensity, recorded in rapidly deposited sediments or lava flow sequences, provide constraints on the dynamics of convection in Earth's outer core.

Together, these observations help geophysicists test and refine numerical models of the geodynamo.

Magnetic Stratigraphy for Dating and Correlation

Dating Sedimentary Sequences

Magnetostratigraphy exploits the fact that polarity reversals are recorded globally and (essentially) simultaneously, making them powerful time markers.

Here's how it works in practice:

1. Collect oriented samples at closely spaced intervals through a sedimentary section.
2. Measure and demagnetize each sample in the lab to isolate the primary NRM and determine whether it records normal or reversed polarity.
3. Construct a magnetic polarity stratigraphy for the section: a column of normal and reversed polarity zones.
4. Compare this pattern to the geomagnetic polarity timescale (GPTS), which is the calibrated global record of polarity reversals dated using radiometric methods (primarily $^{40}Ar/^{39}Ar$ dating of lava flows and ocean-floor magnetic anomalies).
5. Match the observed polarity pattern to the GPTS to assign ages to the section.

Pattern matching works best when you have at least one independent age tie-point (a radiometric date or a well-dated biostratigraphic datum) to anchor the correlation.

Correlating Geological Events

Because polarity reversals are global, magnetostratigraphy allows you to correlate sedimentary sequences across different basins and even different continents with high precision.

This capability is especially valuable for:

- Dating sequences that lack fossils or volcanic ash layers. Many continental sedimentary sections and deep-sea cores have limited biostratigraphic control, and magnetostratigraphy can fill that gap.

• Establishing the timing of major geological events. Magnetic stratigraphy has been used to precisely date the Cretaceous-Paleogene (K-Pg) boundary (~66 Ma), constrain the duration of the Eocene-Oligocene climate transition (~34 Ma), and bracket the Messinian salinity crisis (~5.96–5.33 Ma) when the Mediterranean Sea largely evaporated.

• Building integrated timescales. Magnetostratigraphy is routinely combined with biostratigraphy, cyclostratigraphy (orbital tuning), and radiometric dating to produce the most accurate geological timescales available.

The resolution of magnetostratigraphy depends on the reversal frequency during the time interval of interest. During periods of frequent reversals (like much of the Cenozoic), resolution can be better than 100 kyr. During long intervals of stable polarity (like the Cretaceous Normal Superchron, ~84–124 Ma), magnetostratigraphy provides no age resolution at all.

Unit 5 – Electrical and Electromagnetic Methods

5.1 Electrical properties of rocks and minerals

Electrical Properties of Rocks and Minerals

Electrical properties of rocks and minerals govern how Earth materials interact with electric currents and electromagnetic fields. Understanding conductivity, resistivity, and dielectric permittivity is essential for interpreting geophysical survey data and for resource exploration targeting oil, gas, groundwater, and mineral deposits.

Conductivity and Resistivity

Electrical conductivity (σ) measures how easily a material conducts electric current. Electrical resistivity (ρ) measures how strongly a material opposes current flow. These two properties are inversely related:

$\rho = \frac{1}{\sigma}$

Most rocks are poor conductors (high resistivity), while many ore-forming minerals are good conductors (low resistivity). The range of electrical resistivity in geological materials spans many orders of magnitude, from about 10^{-5} Ω·m for massive sulfides to over 10^{7} Ω·m for dry crystalline rocks.

As a general pattern:

- Igneous rocks tend to have the highest resistivity due to low porosity and few connected pore spaces.
- Metamorphic rocks fall in between, with resistivity depending on foliation, fracturing, and mineral content.
- Sedimentary rocks typically have the lowest resistivity among common rock types because of their higher porosity and fluid content.

Conductivity and Resistivity of Minerals

Minerals conduct electricity through different mechanisms, and this determines where they fall on the resistivity spectrum.

- Metallic minerals such as native metals (gold, silver, copper) and metallic sulfides (pyrite, chalcopyrite, galena) have very low resistivity and high conductivity. Current flows through free electrons in their crystal lattice, similar to how metals conduct in everyday circuits.
- Non-metallic minerals such as quartz, feldspars, and calcite have high resistivity and very low conductivity. They lack free charge carriers and behave as electrical insulators.
- Clay minerals deserve special attention. Their small particle size and high surface area create conditions for surface conduction, where ions adsorbed onto mineral surfaces can move under an applied electric field. Ion exchange at clay surfaces further enhances this effect. The result is that even modest clay content can significantly lower a rock's bulk resistivity, which complicates the interpretation of electrical and electromagnetic survey data because the low resistivity might be mistaken for the presence of saline fluids or ore minerals.

Factors Influencing Electrical Properties

Composition and Porosity

Mineral composition sets the baseline electrical character of a rock. A higher proportion of conductive minerals (e.g., sulfides, graphite) lowers the bulk resistivity.

Porosity refers to the fraction of a rock's volume occupied by open spaces. Its effect on electrical properties depends on three things:

- Volume of pore space — higher porosity generally means lower resistivity, since pore fluids are usually more conductive than the rock matrix.
- Distribution of pores — evenly distributed pores have a different effect than isolated vugs or fractures.
- Connectivity of pore spaces — well-connected pores allow continuous current pathways through the fluid, dramatically lowering resistivity. Isolated pores contribute much less.

Fluid Properties

Pore fluids are often the dominant control on a rock's bulk electrical properties.

- Fluid content: Dry rocks are highly resistive. Adding fluid to the pore space creates ionic conduction pathways that lower resistivity substantially.
- Salinity: Dissolved salts dissociate into ions that carry electric current. Seawater (salinity ~35 g/L, resistivity ~0.2 Ω·m) is far more conductive than freshwater (resistivity ~10–100 Ω·m). Higher salinity means more charge carriers and lower resistivity.
- Temperature: Higher temperatures increase ion mobility in pore fluids, raising conductivity and lowering resistivity. This is relevant in geothermal systems, where elevated subsurface temperatures produce anomalously low resistivity zones that can be mapped with electromagnetic surveys.

Pressure Effects

Increasing confining pressure (e.g., with burial depth) compacts pore spaces and closes microfractures. This reduces porosity and fluid connectivity, which raises resistivity. The effect is most pronounced in rocks with high initial porosity, such as poorly consolidated sedimentary rocks.

In rarer cases, extreme pressures can trigger mineral phase transitions that alter electrical properties. A classic example is the transition from graphite (electrically conductive) to diamond (an insulator) at very high pressures and temperatures. While this specific transition occurs deep in the mantle, it illustrates how pressure-induced structural changes in minerals can fundamentally change their electrical behavior.

Role of Pore Fluids in Electrical Properties

Fluid Conductivity

For most rocks encountered in geophysical surveys, the electrical conductivity of the bulk rock is controlled primarily by its pore fluids rather than by the rock matrix. The mineral grains themselves are typically insulators (quartz, feldspar, calcite), so current flows through the interconnected fluid-filled pore network. This is especially true for porous, permeable sedimentary rocks.

This principle is what makes electrical and electromagnetic methods so effective for hydrogeological and petroleum exploration: you're largely measuring the fluid, not the rock.

Archie's Law

Archie's Law is the foundational empirical relationship linking a rock's bulk conductivity to its porosity, fluid saturation, and pore fluid conductivity:

$$\sigma_r = \sigma_f \cdot \phi^m \cdot S_w^n$$

where:

- σ_r = bulk rock conductivity
- σ_f = pore fluid conductivity
- ϕ = porosity (fractional)
- S_w = water saturation (fraction of pore space filled with water)
- m = cementation exponent (typically 1.3–2.5, reflects pore geometry and connectivity)
- n = saturation exponent (typically ~2)

The cementation exponent m increases with more tortuous, less-connected pore networks. The saturation exponent n captures how partial saturation reduces the available conduction pathways.

Archie's Law is widely used in the oil and gas industry to estimate hydrocarbon saturation from well-log resistivity data. The logic is straightforward: if you know σ_f, ϕ, and m from other measurements, you can solve for S_w. Low water saturation implies the remaining pore space is filled with hydrocarbons.

Limitations: Archie's Law assumes that all electrical conduction occurs through the pore fluid. It breaks down when conductive minerals (sulfides, graphite) or clay minerals contribute significant surface conduction. Deviations from Archie's Law predictions can therefore indicate clay content or the presence of metallic minerals.

Dielectric Permittivity in Geophysical Exploration

Principles of Dielectric Permittivity

Dielectric permittivity (ε) measures a material's ability to store electromagnetic energy when subjected to an external electric field. It controls how electromagnetic waves propagate through a medium, affecting both wave velocity and attenuation.

Relative permittivity (ε_r) expresses a material's permittivity as a ratio to the permittivity of free space (ε_0):

$$\varepsilon_r = \frac{\varepsilon}{\varepsilon_0}$$

Typical values for geological materials:

- Water: $\varepsilon_r \approx 80$
- Most rock-forming minerals: $\varepsilon_r \approx 3\text{–}10$
- Air: $\varepsilon_r \approx 1$

Because water's relative permittivity is so much higher than that of mineral grains or air, even small changes in water content strongly affect the bulk dielectric properties of rocks and soils. Contrasts in dielectric permittivity at subsurface boundaries cause reflections and refractions of electromagnetic waves, which form the basis for several geophysical imaging techniques.

Applications in Geophysical Exploration

Ground-penetrating radar (GPR) is the most direct application of dielectric permittivity contrasts. GPR transmits high-frequency electromagnetic pulses (typically 10 MHz to 2 GHz) into the ground and records reflections from interfaces where ε_r changes. The travel time of reflections gives depth information, while reflection amplitude relates to the magnitude of the permittivity contrast.

Practical applications include:

- Groundwater exploration: Detecting the water table, where a sharp increase in ε_r occurs as pores become saturated.
- Soil moisture monitoring: Bulk ε_r is highly sensitive to volumetric water content, making dielectric measurements a standard tool in environmental and agricultural geophysics.
- Detection of subsurface voids or contaminants: Air-filled cavities (low ε_r) or fluid contaminant plumes (variable ε_r) create detectable contrasts against the surrounding soil or rock.

The bulk dielectric permittivity of a rock or soil depends on mineral composition, porosity, fluid content, and the frequency of the electromagnetic signal. At higher frequencies, dielectric relaxation effects become important, and permittivity can become frequency-dependent. Accounting for these factors is essential when designing surveys and interpreting electromagnetic geophysical data.

5.2 Resistivity and induced polarization methods

Electrical Resistivity Surveys

Principles and Factors Affecting Resistivity

Electrical resistivity surveys work by injecting a known current into the ground through electrodes, then measuring the resulting voltage (potential difference) at separate electrodes. From these measurements, you can calculate how strongly the subsurface material resists the flow of electric current.

Resistivity is the intrinsic property that quantifies this resistance, measured in ohm-meters ($\Omega \cdot m$). It's the inverse of electrical conductivity. The apparent resistivity you measure in the field represents a bulk average over the volume of earth the current passes through, not the true resistivity of any single layer.

Several factors control the resistivity of subsurface materials:

- Rock type — Igneous and metamorphic rocks tend to have high resistivities (often 10^3 to 10^5 $\Omega \cdot m$), while saturated sedimentary rocks are typically much lower (1 to 10^3 $\Omega \cdot m$).
- Porosity and saturation — In most crustal rocks, current flows primarily through pore fluids rather than through mineral grains. Higher porosity and higher saturation mean more conductive pathways, lowering bulk resistivity. This relationship is formalized by Archie's Law: $\rho = a\, \rho_w\, \phi^{-m}\, S_w^{-n}$, where ρ_w is the fluid resistivity, ϕ is porosity, S_w is water saturation, and a, m, n are empirical constants.
- Fluid chemistry — Saline pore water is far more conductive than fresh water. A sandstone saturated with brine might have a resistivity orders of magnitude lower than the same sandstone filled with fresh water.
- Temperature — Higher temperatures increase ion mobility in pore fluids, which decreases resistivity.
- Clay content — Clay minerals contribute additional surface conductivity through their cation exchange capacity, lowering bulk resistivity independent of pore fluid salinity.

Applications and Detection Capabilities

Because different materials and conditions produce distinct resistivity signatures, these surveys can detect a wide range of subsurface features:

- Groundwater exploration — Freshwater-saturated aquifers typically show moderate resistivity (contrasting with surrounding dry rock or clay), while saline aquifers appear as low-resistivity zones.
- Environmental site assessment — Conductive contaminant plumes (e.g., landfill leachate, saltwater intrusion) show up as low-resistivity anomalies against a higher-resistivity background.
- Geotechnical investigations — Mapping bedrock depth, identifying clay-rich zones, and characterizing soil variability for engineering projects.
- Mineral exploration — Massive sulfide ore bodies (pyrite, chalcopyrite, galena) are highly conductive and produce strong low-resistivity anomalies.

Note that resistivity alone doesn't uniquely identify a material. A low-resistivity anomaly could be clay, saline water, or sulfide mineralization. That ambiguity is one reason IP measurements and integration with other data are so important.

Electrode Configurations

Common Electrode Arrays

A standard resistivity measurement uses four electrodes: two current electrodes (commonly labeled A and B, or C1 and C2) that inject and collect current, and two potential electrodes (M and N, or P1 and P2) that measure the resulting voltage. The geometric factor K depends on the specific arrangement and spacing of these four electrodes, and it converts your voltage and current readings into apparent resistivity:

$\rho_a = K\frac{\Delta V}{I}$

The choice of array affects your depth of investigation, spatial resolution, sensitivity to lateral versus vertical structures, and signal strength. The four most common configurations are the Wenner, Schlumberger, dipole-dipole, and pole-pole arrays.

Characteristics and Applications of Electrode Arrays

Wenner array — All four electrodes are equally spaced at distance a. The current electrodes are on the outside, potential electrodes on the inside. The geometric factor is $K = 2\pi a$.

- Strong signal strength (large voltage for a given current), so it's relatively robust against noise.
- Good sensitivity to vertical resistivity changes (horizontal layering).
- Limited sensitivity to lateral variations.
- To increase depth of investigation, you expand the entire array, which means moving all four electrodes for each reading. This makes 2D profiling slower compared to some other arrays.

Schlumberger array — Four collinear electrodes, but the current electrode spacing (AB) is much larger than the potential electrode spacing (MN). The geometric factor is $K = \pi\frac{(AB/2)^2 - (MN/2)^2}{MN}$, which simplifies to $K \approx \pi\frac{L^2}{MN}$ when $AB \gg MN$.

- Greater depth of investigation than Wenner for the same overall spread, because you expand only the current electrodes while keeping the potential electrodes fixed (until the signal drops too low).

- Efficient for vertical electrical sounding (VES) to determine layered resistivity structure.
- Sensitive to vertical variations; less sensitive to lateral heterogeneity near the potential electrodes.

Dipole-dipole array — Two closely spaced current electrodes (separation a) and two closely spaced potential electrodes (also separation a), separated by a distance $n \times a$. The geometric factor is $K = \pi n(n+1)(n+2)\,a$.

- Excellent sensitivity to lateral resistivity changes, making it well suited for mapping vertical structures (faults, dikes, lateral boundaries).
- Signal strength drops rapidly as n increases (proportional to n^{-3}), so it becomes noisy at large electrode separations.
- Very efficient for multi-channel 2D profiling because you can use a fixed cable with many electrodes and switch between combinations electronically.

Pole-pole array — Only one current electrode and one potential electrode are used in the survey area; the return current electrode and reference potential electrode are placed at effectively infinite distance (typically $>$ 10 times the maximum survey electrode spacing).

- Deepest penetration for a given electrode separation and the simplest geometric factor: $K = 2\pi a$.
- Lowest spatial resolution of the common arrays.
- Practically difficult because the remote electrodes must be very far away, and long wire runs pick up noise. Rarely used in practice except in specialized situations.

Quick comparison: Wenner and Schlumberger are best for layered (1D) targets. Dipole-dipole excels at resolving lateral contrasts. In modern 2D/3D surveys, multi-electrode systems often collect data in several array configurations simultaneously to combine their strengths.

Induced Polarization for Exploration

Principles and Measurements

Induced polarization (IP) measures the ability of subsurface materials to store electrical charge temporarily. When you inject current into the ground, charge accumulates at interfaces between mineral grains and pore fluid. After you shut off the current, this stored charge dissipates, producing a decaying secondary voltage. That decay is the IP effect.

Two physical mechanisms produce IP responses:

- Electrode (metallic) polarization — Occurs at the interface between electronically conducting minerals (sulfides, graphite, some oxides) and ionically conducting pore fluid. Charge transfer across this interface is slow, causing strong polarization. This is the dominant mechanism in mineral exploration targets.
- Membrane polarization — Occurs in clay-rich sediments and rocks where narrow pore throats and charged clay surfaces restrict ion flow, creating zones of charge buildup. This produces a weaker but still measurable IP effect.

IP can be measured in two ways:

1. Time-domain IP — You inject a current pulse, switch it off, and measure the decay of the secondary voltage over time. Chargeability (M) is defined as the area under the decay curve normalized by the primary voltage:

$$M = \frac{1}{V_p}\int_{t_1}^{t_2} V_s(t)\,dt$$

Chargeability is reported in milliseconds (ms). Typical values range from a few ms for barren rock to tens or hundreds of ms for disseminated sulfides.

1. Frequency-domain IP — You measure apparent resistivity at two or more frequencies. Because polarizable materials show frequency-dependent resistivity, the percent frequency effect (PFE) or metal factor quantifies the IP response:

$$PFE = \frac{\rho_{DC} - \rho_{AC}}{\rho_{AC}} \times 100\%$$

where ρ_{DC} is the low-frequency resistivity and ρ_{AC} is the high-frequency resistivity.

Applications in Mineral Exploration

IP is one of the most effective geophysical methods for detecting disseminated sulfide mineralization, which resistivity alone often misses. A rock with only 2–5% disseminated sulfide grains may not produce a significant resistivity anomaly, but it can generate a strong chargeability anomaly because each grain acts as a small capacitor.

Key exploration targets for IP surveys:

- Porphyry copper deposits — Disseminated chalcopyrite and pyrite in large, low-grade ore bodies produce broad, high-chargeability anomalies.
- Disseminated gold deposits — Gold itself isn't polarizable, but it's commonly associated with pyrite and arsenopyrite, which are. IP effectively maps the sulfide halo around gold mineralization.
- Volcanogenic massive sulfide (VMS) deposits — Both the massive sulfide core and the surrounding disseminated sulfide alteration zone produce IP anomalies.

A practical caution: graphite and barren pyrite also produce strong IP responses. High chargeability doesn't automatically mean economic mineralization. You need to interpret IP results alongside resistivity data, geological context, and geochemistry.

Interpreting Resistivity and Induced Polarization Data

Data Representation and Inversion

Raw field measurements give you apparent resistivity and apparent chargeability values, which are weighted averages influenced by everything the current passes through. Converting these into a model of the true subsurface property distribution requires several steps:

1.

Pseudosection plotting — Measured values are plotted at a position midway between the electrode array and at a depth proportional to the electrode separation. This creates a 2D cross-section that gives a rough visual impression of subsurface structure, but it is not a true depth section. Pseudosections distort the shapes and positions of anomalies.

2.

Data quality control — Before inversion, noisy or erroneous data points (from electrode contact problems, cultural interference, etc.) are identified and removed.

3.

Inversion — An algorithm iteratively adjusts a gridded subsurface model until the synthetic data it predicts matches the observed data within an acceptable misfit. Common approaches include: - Least-squares (smooth) inversion — Produces a smooth model that minimizes both data misfit and model roughness. Good for broad anomalies but tends to smear sharp boundaries. - Robust (L1-norm) inversion — Allows sharper contrasts in the model, better for resolving discrete boundaries like faults or contacts.

4.

Model assessment — Check the data misfit (RMS error), inspect for inversion artifacts, and evaluate whether the model features are required by the data or are artifacts of the regularization.

The output is a 2D section or 3D volume of true resistivity and chargeability, which you can then interpret geologically.

Integration and Interpretation

Interpreting resistivity and IP models means linking electrical property variations to geology. Some general associations:

- Low resistivity + low chargeability — Clay, saline groundwater, or weathered rock.
- Low resistivity + high chargeability — Sulfide mineralization (the classic exploration target).
- High resistivity + low chargeability — Fresh, competent rock with low porosity; or dry, clean sand/gravel.
- High resistivity + high chargeability — Less common; can indicate disseminated sulfides in a resistive host rock, or sometimes graphitic schist.

These are guidelines, not rules. The same electrical signature can have different geological causes in different settings, which is why integration with other data is essential:

- Borehole logs provide ground-truth measurements of lithology, mineralization, and physical properties at specific locations, allowing you to calibrate your geophysical models.
- Seismic reflection/refraction data reveal structural and stratigraphic geometry that helps constrain the shapes of resistivity anomalies.
- Geological and geochemical mapping provides the regional framework for understanding what's geologically plausible.
- Magnetic and gravity data can help distinguish between different sources of similar resistivity anomalies (e.g., a magnetic high coinciding with a chargeability high strengthens the case for sulfide mineralization over graphite).

Combining resistivity and IP with complementary datasets reduces ambiguity and leads to more reliable geological interpretations.

5.3 Electromagnetic methods and their applications

Electromagnetic Induction Principles

Electromagnetic (EM) methods exploit a simple but powerful idea: a changing magnetic field will induce electric currents in any nearby conductor. In geophysics, the "conductor" is the ground itself. By transmitting a controlled EM field and measuring how the subsurface responds, you can map variations in electrical conductivity at depth. These methods are workhorses for mineral exploration, groundwater mapping, and environmental site characterization.

Faraday's Law and Electromagnetic Induction

Electromagnetic induction is the process by which a changing magnetic field drives an electric current in a conductor. Faraday's law quantifies this: the induced electromotive force (emf) is proportional to the rate of change of magnetic flux through the conductor.

$$\mathcal{E} = -\frac{d\Phi}{dt}$$

where $\mathcal{E}$ is the induced emf and Φ is the magnetic flux. The negative sign reflects Lenz's law: the induced emf always acts to oppose the flux change that created it. This opposition is what generates the secondary fields you actually measure in a survey.

Electromagnetic Induction in Geophysical Surveys

In practice, a transmitter (typically a loop of wire or a long grounded wire) generates a primary EM field. When this time-varying field encounters conductive material underground, it induces circulating eddy currents in that material. Those eddy currents, in turn, produce their own secondary EM field.

A receiver measures the total field, which is the sum of the primary and secondary fields. The secondary field is typically much weaker than the primary, so sensitive instrumentation and careful signal processing are needed to extract it. The amplitude, phase, and frequency content of the secondary field all carry information about subsurface conductivity structure.

Depth of investigation depends on two factors:

- Frequency of the primary field: Lower frequencies penetrate deeper because they diffuse farther before attenuating.
- Subsurface conductivity: More resistive ground allows deeper penetration; highly conductive ground attenuates the signal quickly, limiting depth.

This frequency-conductivity tradeoff is governed by the skin depth concept, which describes how far an EM wave penetrates before its amplitude drops to $1/e$ (about 37%) of its surface value.

Frequency vs. Time-Domain Methods

The two main families of EM methods differ in how they transmit and record the signal. Each has distinct advantages depending on your target depth, resolution needs, and survey logistics.

Frequency-Domain Electromagnetic (FDEM) Methods

FDEM methods transmit a continuous sinusoidal primary field at one or more fixed frequencies. Common systems include ground conductivity meters (e.g., Geonics EM31, EM34) and airborne platforms (e.g., DIGHEM, RESOLVE). Operating frequencies typically range from a few hundred Hz to a few hundred kHz.

The receiver measures the secondary field in terms of two quantities relative to the primary:

- Amplitude: proportional to the conductivity of the subsurface.
- Phase shift: related to the ratio of conductive to resistive properties.

The secondary field is decomposed into two components:

- In-phase (real) component: influenced by both conductivity and magnetic susceptibility. This makes it useful for identifying magnetic materials but also means it can be harder to interpret for conductivity alone.

- Quadrature (imaginary) component: primarily sensitive to conductivity, making it the go-to component for conductivity mapping.

FDEM systems are generally compact, fast, and well-suited for shallow investigations (typically the upper tens of meters, depending on frequency and coil separation).

Time-Domain Electromagnetic (TDEM) Methods

TDEM methods take a different approach. Instead of a continuous sinusoid, the transmitter drives a current pulse through a loop, then abruptly shuts it off. Common systems include the TEM47, NanoTEM, and the airborne VTEM. The primary pulse is typically a square wave lasting milliseconds to seconds.

After the transmitter switches off, the receiver records the decay of the secondary field during the off-time. The decay rate directly reflects subsurface conductivity:

- Conductive materials sustain eddy currents longer, producing a slow decay.
- Resistive materials lose their eddy currents quickly, producing a fast decay.

Because you're measuring during the off-time (when the primary field is absent), TDEM avoids the challenge of separating a weak secondary field from a strong primary. This is a major practical advantage.

TDEM is also less affected by magnetic susceptibility than FDEM, since the off-time response is dominated by conductivity. Depth of investigation is generally greater than FDEM, ranging from a few meters to several hundred meters depending on the transmitter moment (current × loop area) and ground conductivity.

FDEM vs. TDEM at a glance: - FDEM: continuous signal, measures amplitude and phase, good for shallow targets, sensitive to both conductivity and magnetic susceptibility - TDEM: pulsed signal, measures decay curve, better depth penetration, primarily sensitive to conductivity

Electromagnetic Applications in Geoscience

Mineral Exploration

EM methods are among the most effective tools for detecting conductive ore bodies. The key targets include:

- Massive sulfide deposits (e.g., volcanogenic massive sulfides): highly conductive because interconnected sulfide minerals form continuous current paths.
- Graphite deposits: conductive due to the inherent electrical properties of graphite.
- Nickel-copper sulfide deposits (e.g., magmatic sulfides): conductive for the same reason as massive sulfides, with interconnected sulfide networks.

Airborne EM (AEM) surveys are the standard for large-scale reconnaissance. They cover vast areas quickly and cost-effectively, identifying anomalies that warrant follow-up. AEM data are typically integrated with magnetic and radiometric data to build a more complete geological picture.

Ground-based EM surveys provide higher resolution for detailed target characterization:

- Moving-loop methods (e.g., using SQUID sensors or SMARTem receivers): the transmitter and receiver move together along survey lines, producing detailed conductivity profiles.
- Fixed-loop methods (e.g., UTEM, DEEPEM): a large stationary transmitter loop energizes the ground while one or more receivers map the response. Fixed-loop setups generate stronger signals and are

better for detecting deep or weakly conductive targets.

Environmental Studies

EM methods are valuable in environmental work because they're non-invasive and can cover large areas rapidly.

- Contaminant plume mapping: Conductive plumes from leaking underground storage tanks or landfill leachate show up clearly against a more resistive background.
- Buried object detection: Metal drums, pipes, and other buried waste produce strong EM anomalies.
- Geological characterization: Clay layers, aquitards, and other lithological boundaries can be identified from their conductivity contrasts, which is critical for understanding groundwater flow.

Borehole EM methods add vertical resolution to surface surveys:

- Downhole conductivity logging provides a detailed conductivity profile along a borehole, useful for identifying lithological boundaries and contaminated zones.
- Cross-hole tomography measures EM responses between boreholes to produce 2D or 3D images of the conductivity distribution between wells.

Interpreting Electromagnetic Data

Data Presentation and Interpretation

EM data are displayed in three main formats, each suited to different interpretation goals:

- Profiles: show the EM response along a survey line. Good for spotting anomalies and lateral conductivity changes.
- Maps: show the spatial distribution of the EM response across the survey area. Useful for identifying conductive zones and structural trends.
- Depth sections: show how conductivity varies with depth. These reveal the vertical distribution of conductive layers and targets.

Conductive targets appear as anomalies with higher amplitude and/or a more pronounced phase shift relative to the background. The anomaly shape tells you about the target geometry:

- Narrow, high-amplitude anomalies typically indicate shallow, compact conductors.
- Broad, low-amplitude anomalies suggest deeper or more laterally extensive targets.
- The orientation of an anomaly can reveal the strike and dip of the conductive body.

Advanced Interpretation Techniques

For FDEM data, the quadrature and in-phase components serve different roles. The quadrature component is your primary tool for mapping conductivity structure. The in-phase component helps identify magnetic materials and estimate magnetic susceptibility, but its sensitivity to both properties can complicate interpretation.

For TDEM data, the decay curve contains depth information:

- Early-time measurements (shortly after transmitter shutoff) are sensitive to shallow conductors.

- Late-time measurements sample progressively deeper, as the eddy current system diffuses downward through the subsurface over time.

Forward modeling and inversion are the standard tools for extracting quantitative conductivity models from EM data:

1. Forward modeling: You define a conductivity model, calculate the EM response it would produce given your survey geometry, and compare the predicted response to your measured data.

2. Inversion: The conductivity model is adjusted iteratively to minimize the misfit between calculated and observed data. The algorithm searches for the model that best explains your measurements.

3. Common inversion approaches include least-squares, regularized (which adds smoothness or other constraints to stabilize the solution), and smooth-model inversions.

Inversion is never unique. Multiple conductivity models can fit the same data, so geological constraints and independent information (e.g., from drilling) are essential for producing reliable interpretations.

5.4 Ground-penetrating radar and its applications

Ground-Penetrating Radar Principles

GPR System Components and Operation

Ground-penetrating radar (GPR) is a geophysical method that sends high-frequency electromagnetic (EM) pulses into the ground and listens for reflections. Those reflections come from interfaces between materials with different electrical properties, giving you an image of what's buried below without ever breaking the surface.

A GPR system has two main parts:

- Transmitter: Emits short pulses of EM energy, typically in the 10 MHz to 2 GHz range (MHz = megahertz, GHz = gigahertz)

- Receiver: Records the amplitude and travel time of reflected signals returning from subsurface interfaces and objects

The receiver output is processed and displayed as a radargram (also called a GPR profile), which is essentially a cross-sectional image of the subsurface along your survey line.

EM waves propagate through earth materials at velocities controlled by three properties: dielectric permittivity, magnetic permeability, and electrical conductivity. Of these, dielectric permittivity is usually the dominant factor.

Reflection and Propagation of GPR Signals

Reflections occur wherever there's a contrast in dielectric properties. Think of the boundary between dry sand and a clay layer, or between soil and a buried pipe. The greater the contrast, the stronger the reflection.

Dielectric permittivity controls wave velocity through a material. Higher permittivity means slower propagation. You can estimate the velocity with:

$$v = \frac{c}{\sqrt{\varepsilon_r}}$$

where v is the wave velocity in the material, c is the speed of light in a vacuum ($\approx 3 \times 10^8$ m/s), and ε_r is the relative dielectric permittivity.

For example, dry sand has $\varepsilon_r \approx 3\text{–}5$, giving velocities around 0.13–0.17 m/ns. Saturated clay might have $\varepsilon_r \approx 15\text{–}40$, slowing the wave considerably.

The two-way travel time (the time for a pulse to travel down to a reflector and back) is what you measure. If you know the velocity, you can convert that travel time into depth:

$d = \frac{v \cdot t}{2}$

where d is depth and t is two-way travel time.

GPR Penetration and Resolution

Factors Affecting Penetration Depth

There's a fundamental trade-off in GPR: lower frequencies penetrate deeper but resolve less detail, while higher frequencies give finer resolution but die out quickly.

- Frequencies in the 10–500 MHz range can reach depths of tens of meters in favorable materials, but they won't resolve small features.
- Frequencies in the 500 MHz–2 GHz range may only penetrate centimeters to a few meters, but they can pick up fine details like individual rebar in concrete.

The electrical conductivity of subsurface materials is the biggest enemy of penetration depth. Conductive materials absorb EM energy rapidly. In practice:

- Good for GPR (resistive): dry sand, gravel, limestone, granite, ice. These allow deep penetration.
- Bad for GPR (conductive): clay-rich soils, saltwater-saturated sediments, shale. These attenuate the signal quickly, sometimes limiting penetration to less than a meter.

Water content matters too, because water has a high dielectric permittivity ($\varepsilon_r \approx 80$). Increasing moisture raises the bulk permittivity, slows the wave, and generally increases attenuation.

Vertical and Horizontal Resolution

Vertical resolution is your ability to distinguish two closely spaced layers. It's approximately one-quarter of the wavelength:

$\Delta z \approx \frac{\lambda}{4} = \frac{v}{4f}$

where f is the antenna frequency and v is the wave velocity. A 400 MHz antenna in dry sand ($v \approx 0.15$ m/ns) gives a vertical resolution of roughly 0.09 m, or about 9 cm. Features thinner than this blend together and can't be distinguished.

Horizontal resolution depends on the antenna footprint, which is the cone of energy illuminating the subsurface at any given point. It's governed by:

- Antenna frequency (higher = tighter footprint)
- Depth to the target (deeper targets have larger footprints)
- Velocity of the medium

The Fresnel zone defines the area on a reflector that contributes energy back to the receiver. Smaller Fresnel zones mean better horizontal resolution. This is one reason migration processing (discussed below) is so valuable: it effectively shrinks the Fresnel zone in your final image.

Applications of GPR

Utility Detection and Mapping

GPR is one of the most common tools for locating buried utilities (pipes, cables, storage tanks) before excavation or construction. Hitting an unknown gas line or fiber-optic cable is expensive and dangerous, so accurate mapping is critical.

GPR can often differentiate metallic from non-metallic utilities. Metal objects produce strong, high-amplitude hyperbolic reflections because of the large dielectric contrast with surrounding soil. Non-metallic utilities (PVC pipes, concrete conduits) produce weaker reflections but are still detectable when conditions are favorable.

Archaeological and Forensic Investigations

GPR lets archaeologists map buried structures, foundations, walls, and artifacts without excavation. This is especially valuable at sensitive heritage sites where destructive sampling must be minimized. Survey grids can produce 3D maps of an entire site, guiding targeted excavation to the most promising areas.

In forensic work, GPR is used to locate unmarked graves, clandestine burials, and hidden objects. Disturbed soil has different dielectric properties than undisturbed ground, and decomposition processes create detectable contrasts. These applications require careful interpretation, since many natural features can mimic forensic targets.

Geotechnical and Structural Assessments

GPR is routinely used to evaluate concrete structures, pavements, and bridges:

- Concrete inspection: Locating rebar, measuring slab thickness, and identifying voids, delamination, or areas of deterioration
- Pavement evaluation: Measuring layer thicknesses (asphalt, base, subbase), detecting moisture accumulation or voids beneath the surface
- Bridge decks: Mapping rebar corrosion and concrete degradation to prioritize repair

These surveys help engineers optimize maintenance schedules and direct repair budgets where they're most needed.

Environmental and Geological Applications

- Contamination mapping: Delineating contamination plumes, locating buried waste, and monitoring remediation progress. GPR can detect the boundaries of landfills and identify leaking underground storage tanks.
- Stratigraphy: Imaging bedding planes, depositional structures, and erosional surfaces in the shallow subsurface. This helps reconstruct past depositional environments and assess the geometry of geological units.
- Glaciology: Measuring ice thickness, mapping internal layering, and studying glacier dynamics. Ice is highly resistive, making it an excellent medium for GPR with penetration depths that can reach hundreds of meters.

Interpreting GPR Data

Radargram Analysis and Interpretation

A radargram displays distance along the survey line on the horizontal axis and two-way travel time (or estimated depth) on the vertical axis. Learning to read these profiles is the core skill of GPR interpretation.

Key patterns to recognize:

- Hyperbolic reflections: Point objects (pipes, boulders, rebar) produce characteristic hyperbolas. The apex marks the object's horizontal position, and the shape of the hyperbola depends on wave velocity and object depth.
- Continuous, sub-horizontal reflections: These typically represent stratigraphic interfaces like soil layer boundaries or bedding planes.
- Disrupted or chaotic reflections: May indicate disturbed ground, fill material, or fractured rock.

Reflection polarity carries useful information. A positive polarity (same phase as the transmitted pulse) indicates an increase in dielectric permittivity across the interface, such as going from dry soil into saturated material. A negative polarity (phase reversal) indicates a decrease, such as soil overlying an air-filled void.

Advanced Processing Techniques

Raw GPR data often need processing to become interpretable. Common techniques include:

1. Migration: Collapses hyperbolic reflections back to their true subsurface positions. Algorithms like Kirchhoff or Stolt migration use velocity information to refocus the energy, producing a sharper, more spatially accurate image.
2. Topographic correction: Adjusts for elevation changes along the survey line. Without this, depth estimates on hilly terrain will be distorted.
3. Gain adjustments: Compensate for signal attenuation with depth so that deeper reflections are visible alongside shallow ones.
4. Amplitude analysis and attribute extraction: Highlight specific subsurface properties, such as changes in material type or the presence of fluids.
5. 3D rendering and time-slice maps: When data are collected on a grid, horizontal slices at specific depths can reveal the plan-view shape of buried features. This is especially powerful in archaeological and utility surveys.

Data Interpretation Considerations

GPR data never exist in a vacuum. Good interpretation requires integrating several sources of information:

- Site context: Knowing the local geology, site history, and existing infrastructure helps you distinguish real targets from noise. A hyperbola in a radargram could be a buried pipe or a tree root; site knowledge helps you decide.
- Noise sources: External interference from radio transmitters, power lines, or nearby vehicles can contaminate data. Surface objects like fences or parked cars can also generate reflections that appear to come from the subsurface.

- Complementary data: Borehole logs, geotechnical reports, and results from other geophysical methods (resistivity, magnetics) should be used to validate GPR interpretations. No single method gives the full picture.

Effective GPR projects typically involve collaboration between geophysicists, geologists, engineers, and site stakeholders. The geophysicist processes and interprets the data, but domain experts provide the context needed to turn a radargram into actionable information.

Unit 6 – Heat Flow and Geothermal Energy

6.1 Earth's heat budget and heat flow

Heat Distribution within Earth's Interior

Earth's heat budget describes where the planet's internal thermal energy comes from and how it moves from the deep interior toward the surface. This topic matters because heat flow is the engine behind plate tectonics, volcanism, and the geodynamo that generates Earth's magnetic field. Understanding it connects nearly every other concept in geophysics.

Sources of Earth's Internal Heat

Earth's internal heat comes from two main sources: radioactive decay and primordial heat.

Radioactive decay is the dominant source, accounting for roughly 80% of the total internal heat budget. Unstable isotopes of uranium (^{238}U, ^{235}U), thorium (^{232}Th), and potassium (^{40}K) undergo spontaneous nuclear breakdown, releasing energy as heat. These isotopes are concentrated in the crust and upper mantle because they are incompatible elements that preferentially partition into silicate melts during differentiation.

Primordial heat makes up the remaining ~20%. This is leftover thermal energy from:

- The accretion of planetesimals during Earth's formation
- Gravitational energy released during core-mantle differentiation
- Giant impacts (most notably the Moon-forming impact)

Primordial heat has been slowly dissipating over 4.5 billion years, but Earth is large enough that significant amounts remain stored in the deep interior.

Uneven Distribution of Heat

Heat production and temperature are distributed unevenly through Earth's layers, and it's worth keeping the distinction between the two clear.

Heat production (the rate at which new thermal energy is generated) is highest in the crust and upper mantle, where radioactive elements are most concentrated. The lower mantle and core contain far fewer radioactive isotopes, so they produce less heat per unit volume.

Temperature, on the other hand, increases with depth. Earth's core reaches an estimated 5,000–6,000°C. Those extreme temperatures reflect:

- Residual primordial heat trapped at depth
- The release of latent heat as the liquid outer core gradually crystallizes onto the solid inner core
- The insulating effect of the overlying mantle, which slows heat loss

So the crust produces more heat per kilogram, but the core is far hotter. This distinction trips people up on exams.

Heat Transfer Mechanisms in Earth

Three mechanisms move heat through the planet: conduction, convection, and radiation. Each dominates in different regions.

Conduction in the Lithosphere

Conduction is the transfer of thermal energy through direct atomic or molecular contact, with no bulk movement of material. One atom vibrates, transfers kinetic energy to its neighbor, and so on. This is the primary way heat moves through the rigid lithosphere (crust and uppermost mantle).

The rate of conductive heat transfer depends on thermal conductivity (k), which varies by rock type and mineral composition:

- Quartz-rich rocks (e.g., quartzite) conduct heat relatively well
- Feldspar-rich rocks conduct heat more slowly
- Olivine has higher thermal conductivity than pyroxene

Because conduction is slow compared to convection, the lithosphere acts as a thermal lid on the planet, controlling how efficiently Earth loses heat to the surface.

Convection in the Mantle

Convection is the dominant heat transfer mechanism in the mantle and is the process most directly responsible for plate tectonics. It involves the bulk movement of material driven by density differences: hotter, less dense rock rises while cooler, denser rock sinks.

How mantle convection works, step by step:

1. Heat from the core and from radioactive decay warms the lower mantle
2. The heated rock becomes less dense and rises buoyantly
3. As it approaches the upper mantle and lithosphere, it spreads laterally and loses heat
4. The cooled rock becomes denser and eventually sinks back down
5. This cycle forms large-scale convection cells spanning thousands of kilometers

These convection cells interact with the lithosphere at plate boundaries:

- Divergent boundaries (mid-ocean ridges): Rising limbs of convection push plates apart
- Convergent boundaries (subduction zones): Sinking limbs pull cold lithosphere back into the mantle

The relationship between convection and plate motion is more complex than a simple conveyor belt, but at this level, the connection between rising/sinking mantle flow and plate movement is the key takeaway.

Radiation in the Deep Interior

Radiative heat transfer involves the emission and absorption of electromagnetic energy (primarily infrared). In most of Earth's interior, radiation is a minor contributor compared to conduction and convection. However, it becomes more significant at extreme temperatures because radiative output scales with T^4.

- In the lower mantle, radiative transfer may contribute up to ~30% of the total heat flux

- In the outer core, radiative transfer likely plays a supporting role alongside convection, which generates Earth's magnetic field through the geodynamo

The exact contribution of radiation in the deep Earth remains an active area of research, partly because measuring the radiative properties of minerals at extreme pressures and temperatures is experimentally difficult.

Heat Flow and Tectonic Settings

Surface heat flow, measured in mW/m^2, varies systematically with tectonic setting. These patterns are some of the strongest evidence linking mantle convection to surface geology.

High Heat Flow: Active Tectonic Regions

Regions with active tectonics show elevated heat flow because hot mantle material sits at shallow depths.

- Mid-ocean ridges have the highest heat flow values, typically exceeding $100\ \text{mW/m}^2$. Hot asthenospheric material upwells directly beneath a very thin lithosphere. Heat flow decreases systematically with distance from the ridge axis as the new oceanic crust cools and the lithosphere thickens. This cooling pattern follows a predictable square-root-of-age relationship.

- Continental rift valleys also show elevated heat flow. The East African Rift System, for example, records values of $60\text{–}110\ \text{mW/m}^2$, well above the continental average, because the lithosphere is being stretched and thinned, allowing hot mantle to rise closer to the surface.

- Volcanic arcs above subduction zones display locally high heat flow due to magma generation in the mantle wedge.

Lower Heat Flow: Stable and Subduction Regions

- Subduction zones generally show lower heat flow on the trench side, because cold oceanic lithosphere descends into the mantle, depressing temperatures. The back-arc region behind the trench, however, can show elevated heat flow where the mantle wedge is heated by fluids released from the subducting slab.

- Cratons (old, stable continental interiors like the Canadian Shield or West African Craton) typically have heat flow below $60\ \text{mW/m}^2$. Their thick, cold lithospheric roots extend deep into the mantle, and much of the original radioactive heat production in their crust has decayed over billions of years.

- Old oceanic crust far from ridges also shows low heat flow, reflecting the progressive cooling of the lithosphere with age.

Inferring Subsurface Thermal Conditions

Measuring Heat Flow

Heat flow (Q) is determined from two quantities using Fourier's law of heat conduction:

$Q = k \cdot \frac{dT}{dz}$

where: - Q = heat flow (mW/m^2) - k = thermal conductivity of the rock (W/m·K) - $\frac{dT}{dz}$ = geothermal gradient, the rate of temperature increase with depth ($°\text{C/km}$)

To measure heat flow in practice:

1. Drill a borehole (on land) or deploy a marine heat flow probe (on the seafloor)

2. Measure the temperature at multiple depths to determine the geothermal gradient

3. Collect rock or sediment samples and measure their thermal conductivity in the lab

4. Multiply the gradient by the conductivity to get heat flow

The global average surface heat flow is approximately $87\ \text{mW/m}^2$, but values range from below $40\ \text{mW/m}^2$ in old cratons to over $200\ \text{mW/m}^2$ at mid-ocean ridges.

Interpreting Heat Flow Patterns

Heat flow values tell you about what's happening beneath the surface:

- High values indicate thin lithosphere and/or hot material at shallow depth. Think magma chambers, young oceanic crust, hydrothermal systems, and active volcanic regions.
- Low values indicate thick, cold lithosphere. Think old continental shields and aged oceanic plates.
- Localized anomalies can reveal buried features. A patch of unusually high heat flow in an otherwise stable continental region might indicate a shallow intrusion, an active geothermal system, or elevated crustal radioactivity.

Integrating Heat Flow with Other Geophysical Data

Heat flow measurements become much more powerful when combined with other geophysical observations:

- Seismic data: Seismic wave velocities decrease in hotter, partially molten rock. Combining seismic tomography with heat flow data helps map thermal structure at depth.
- Gravity data: Density variations caused by temperature differences produce gravity anomalies. A region with high heat flow and a negative gravity anomaly likely has hot, low-density material at depth.
- Together, these datasets allow construction of 3D thermal models of the lithosphere and upper mantle, which are essential for understanding mantle dynamics, assessing geothermal energy potential, and interpreting tectonic processes.

6.2 Geothermal gradients and heat flow measurement

Geothermal gradients measure how Earth's temperature changes with depth. This concept is central to understanding heat flow from the planet's interior to its surface, and it directly informs geothermal energy exploration, hydrocarbon maturation studies, and models of Earth's thermal evolution.

Measuring geothermal gradients involves recording temperatures in boreholes and analyzing the thermal properties of the rocks those boreholes penetrate. Combining these two datasets through Fourier's law yields heat flow values that reveal subsurface thermal patterns.

Geothermal Gradient and Its Significance

Definition and Measurement Units

The geothermal gradient is the rate at which Earth's temperature increases with depth. It's typically expressed in °C/km (or sometimes °F/100 ft in older literature). A steeper gradient means temperature rises faster as you go deeper.

Average Gradient and Variability

The average crustal geothermal gradient is roughly 25–30 °C/km, but this number varies a lot depending on location. Tectonic setting, rock type, and subsurface fluid circulation all play a role. In volcanic regions like Iceland, gradients can exceed 100 °C/km, while stable continental interiors may sit well below the average.

Driving Factors and Heat Sources

Two primary sources supply Earth's internal heat:

- Primordial heat from planetary accretion and core formation, left over from Earth's formation ~4.5 billion years ago
- Radiogenic heat from the decay of radioactive isotopes, primarily uranium (U), thorium (Th), and potassium (^{40}K)

Radiogenic heat production is especially significant in felsic crustal rocks. Granitic intrusions, for example, can have elevated heat production rates (on the order of 2–5 μW/m³), which raises local geothermal gradients above the regional background.

Importance in Geophysical Applications

Geothermal gradient data feeds into several applied problems:

- Geothermal energy exploration: identifying high-temperature reservoirs suitable for power generation
- Hydrocarbon maturation studies: determining whether source rocks have reached the thermal window for oil or gas generation
- Tectonic and volcanic analysis: mapping lithospheric thermal structure and identifying active heat sources
- Basin modeling: constraining thermal histories and calibrating numerical simulations of sedimentary basin evolution

Measuring Geothermal Gradients and Heat Flow

Borehole Temperature Measurements

Direct temperature measurement in boreholes is the most reliable way to determine the geothermal gradient. Here's how it works:

1. Drill or access a borehole that penetrates to the depth of interest.
2. Allow the borehole to thermally equilibrate. Drilling circulates fluid that disturbs the natural temperature profile, so the well needs time (often weeks to months) to return to equilibrium.
3. Lower a temperature logging tool (typically a thermistor or thermocouple sensor) down the borehole, recording temperature at regular depth intervals.
4. Plot temperature vs. depth. The slope of this profile gives you the geothermal gradient.

Differential temperature logging is a variant that measures the temperature difference between two closely spaced sensors. This approach is more sensitive to small gradient changes and helps identify interval-specific variations that a single-sensor log might smooth over.

Thermal Conductivity Measurements

Knowing the gradient alone isn't enough to calculate heat flow. You also need to know how easily heat moves through the rock, which is its thermal conductivity (k).

Thermal conductivity is measured in the lab on core samples or cuttings using methods such as:

- Divided bar method: a steady-state technique where a rock disc is sandwiched between materials of known conductivity, and heat flow through the stack is measured
- Needle probe method: a transient technique where a heated needle is inserted into the sample and the rate of temperature rise is recorded

Once you have both gradient and conductivity, you calculate heat flow using Fourier's law of heat conduction:

$Q = -k\frac{dT}{dz}$

- Q = heat flow (mW/m²)
- k = thermal conductivity (W/m·K)
- $\frac{dT}{dz}$ = geothermal gradient (°C/m or K/m)

The negative sign indicates heat flows in the direction of decreasing temperature (from hot to cold, i.e., upward toward the surface).

Estimations from Well Logs

When direct temperature measurements aren't available, geothermal gradients can be estimated indirectly from well logs. Resistivity logs and acoustic (sonic) logs are both sensitive to temperature because temperature affects pore fluid properties and rock matrix behavior. These estimates are less precise than direct measurements but useful when equilibrated temperature logs don't exist.

Heat Flow Measurements

Heat flow values, reported in mW/m², combine gradient and conductivity data into a single quantity that characterizes the thermal regime of a region. Global continental heat flow averages around 65 mW/m², while oceanic heat flow averages about 100 mW/m² (though it varies strongly with crustal age).

Heat flow maps are used to:

- Identify thermal anomalies that may indicate geothermal resources or active tectonic processes
- Constrain regional thermal models
- Compare thermal regimes across different geological provinces

Factors Influencing Geothermal Gradients

Lithology and Thermal Properties

Different rock types conduct heat at different rates, and this directly affects the geothermal gradient. The relationship is inverse: for a given heat flow, rocks with lower thermal conductivity will have higher geothermal gradients, and vice versa.

• Sedimentary rocks (shales, mudstones) generally have low thermal conductivities (1–2 W/m·K), producing steeper gradients in sedimentary basins.

• Salt and quartzite are highly conductive (5–7 W/m·K for salt), so they transmit heat efficiently and produce lower local gradients. Salt diapirs, for instance, act as thermal chimneys, funneling heat upward and creating distinctive thermal anomalies.

• Igneous and metamorphic rocks typically fall between these extremes, though composition matters (felsic rocks tend to be less conductive than mafic rocks).

Tectonic Setting

The tectonic environment exerts first-order control on heat flow and geothermal gradients:

• Extensional settings (continental rifts, back-arc basins) tend to have elevated gradients because lithospheric thinning brings hot asthenospheric material closer to the surface. The Basin and Range Province in the western U.S. is a classic example, with heat flow values commonly exceeding 80–100 mW/m².

• Convergent margins display complex thermal patterns. The forearc above a subducting slab is typically cold (the descending plate depresses isotherms), while the volcanic arc behind it shows elevated heat flow from magmatic activity.

• Stable cratons generally have low, uniform heat flow (40–50 mW/m²) due to their thick, old lithosphere.

Fluid Circulation Effects

Subsurface fluid movement can redistribute heat far more efficiently than conduction alone, and this is one of the biggest complications in interpreting geothermal data:

• Upward fluid migration (e.g., hydrothermal convection) transports heat toward the surface, creating localized hot spots and steepening the near-surface gradient.

• Downward fluid flow (e.g., meteoric water recharge) carries cool surface water to depth, depressing the gradient.

• Convective heat transfer by circulating fluids can dominate over conduction in fractured or permeable zones, making it difficult to extract the purely conductive thermal signal.

Recognizing the signature of advective (fluid-driven) vs. conductive heat transfer is critical when interpreting borehole temperature profiles.

Radiogenic Heat Production

Rocks enriched in U, Th, and ^{40}K generate heat internally through radioactive decay. This effect is most pronounced in:

• Upper-crustal granitic bodies

• Felsic gneisses

• Some sedimentary formations enriched in organic matter or heavy minerals

In regions with thick, radiogenic upper crust, the contribution to surface heat flow can be substantial (sometimes 30–50% of the total).

Interpreting Geothermal Data for Subsurface Characterization

Identifying Thermal Anomalies

Mapping geothermal gradients and heat flow across a region reveals areas that deviate from the background thermal field. Positive anomalies (higher than expected) may indicate:

- Shallow magmatic intrusions or hydrothermal systems
- Geothermal reservoirs suitable for energy extraction
- Upward fluid migration pathways

Negative anomalies can point to downward fluid flow, recent sedimentation (which buries and insulates the surface), or the presence of highly conductive lithologies redistributing heat laterally.

Constraining Basin Thermal History

Geothermal data integrates with other geological and geophysical datasets (vitrinite reflectance, fission track thermochronology, burial history curves) to reconstruct how a sedimentary basin's temperature field has evolved over geologic time. This thermal history is essential for predicting:

- When and where source rocks entered the oil and gas generation windows
- The timing and pathways of hydrocarbon migration
- The degree of diagenetic alteration in reservoir rocks

Estimating Depth to the Brittle-Ductile Transition

The brittle-ductile transition is the depth at which crustal rocks shift from fracturing under stress to flowing plastically. Its depth depends strongly on the geothermal gradient because rock strength decreases with temperature. A steeper gradient pushes this transition shallower, which has implications for:

- Maximum depth of seismicity (earthquakes generally nucleate in the brittle zone)
- Fluid flow pathways (fracture permeability dominates above the transition)
- Crustal deformation style (faulting vs. ductile shear)

Geothermal Resource Assessment

Heat flow and gradient data are the starting point for evaluating geothermal energy potential. High gradients mean you can reach economically useful temperatures (typically >150 °C for electricity generation) at shallower, more drillable depths. Lower-temperature resources (50–100 °C) can still be valuable for direct-use applications like district heating, greenhouse agriculture, and aquaculture.

Calibrating Basin and Thermal Models

Measured geothermal gradients and heat flow values serve as ground truth for numerical models of basin evolution and lithospheric thermal structure. These measurements help:

- Set appropriate thermal boundary conditions (e.g., basal heat flow at the base of the sedimentary section)
- Validate model predictions against observed temperature-depth profiles

- Reduce uncertainty in forward models used for resource exploration and tectonic reconstruction

6.3 Geothermal energy exploration and exploitation

Geothermal energy taps into Earth's heat for power and heating. This renewable resource is found in volcanic areas, plate boundaries, and hot spots. Exploring and exploiting geothermal energy involves identifying high heat flow zones and using advanced technologies.

Geophysical methods like seismic and electromagnetic surveys help locate geothermal reservoirs. Power plants then extract heat to generate electricity or provide direct heating. Environmental impacts are generally low, but initial costs can be high.

Favorable Geological Settings for Geothermal Energy

High Heat Flow Areas

- Geothermal energy resources are typically associated with areas of high heat flow, such as volcanic regions (Yellowstone National Park), tectonic plate boundaries (Ring of Fire), and hot spots (Hawaii)
- The presence of cap rocks, such as impermeable clay layers or dense volcanic rocks, can help to trap and concentrate geothermal fluids in reservoirs, increasing the potential for geothermal energy development

Hydrothermal Systems and Enhanced Geothermal Systems (EGS)

- Hydrothermal systems, which involve the circulation of hot water or steam through permeable rocks, are the most common type of geothermal resource exploited for energy production
- High-temperature hydrothermal systems (>150°C) are often found in volcanic regions and are suitable for electricity generation
- Low-temperature hydrothermal systems (<150°C) are more widespread and are used for direct heating applications, such as space heating (district heating systems) and agricultural processes (greenhouses)
- Enhanced Geothermal Systems (EGS) involve the creation of artificial geothermal reservoirs by fracturing hot, dry rock and circulating fluid through the fractures to extract heat
- Sedimentary basins with high geothermal gradients and thick, permeable aquifers (Williston Basin) can also be potential targets for geothermal energy development

Geophysical Methods in Geothermal Exploration

Seismic, Gravity, and Magnetic Surveys

- Geophysical methods are used to identify and characterize geothermal resources by measuring various physical properties of the subsurface
- Seismic surveys, including reflection and refraction methods, can provide information on subsurface geology, fault structures, and the presence of geothermal reservoirs
- Gravity surveys measure variations in the Earth's gravitational field, which can indicate the presence of low-density, high-temperature rocks and fluids associated with geothermal systems
- Magnetic surveys detect variations in the Earth's magnetic field, which can be influenced by the presence of magnetic minerals altered by geothermal fluids

Electrical, Electromagnetic, and Thermal Methods

- Electrical and electromagnetic methods, such as magnetotellurics (MT) and controlled-source electromagnetic (CSEM) surveys, can map the electrical conductivity of the subsurface, which is often enhanced by the presence of geothermal fluids
- Heat flow measurements and temperature gradient surveys can directly assess the thermal regime of an area and help identify regions with elevated geothermal potential
- Geochemical analysis of surface manifestations (hot springs, fumaroles) can provide insights into the temperature, composition, and origin of geothermal fluids

Principles and Technologies of Geothermal Exploitation

Geothermal Power Generation

- Geothermal energy exploitation involves the extraction of heat from geothermal reservoirs and its conversion into usable forms of energy, such as electricity or direct heat
- Conventional geothermal power plants use steam or hot water from hydrothermal reservoirs to drive turbines and generate electricity
- Flash steam plants are used for high-temperature resources, where the geothermal fluid is rapidly depressurized to produce steam that drives the turbines
- Binary cycle plants are used for lower-temperature resources, where the geothermal fluid heats a secondary working fluid with a lower boiling point (isobutane), which then drives the turbines

Direct Use and Advanced Technologies

- Direct use applications involve the direct utilization of geothermal heat for space heating, greenhouse heating, aquaculture (fish farming), and industrial processes (food dehydration)
- Ground-source heat pumps (GSHPs) use the relatively constant temperature of the shallow subsurface to provide heating and cooling for buildings
- Enhanced Geothermal Systems (EGS) involve the creation of artificial geothermal reservoirs by fracturing hot, dry rock and circulating fluid through the fractures to extract heat
- Hydraulic stimulation techniques, such as hydroshearing and hydraulic fracturing, are used to enhance the permeability of the rock and create a heat exchange system
- Directional drilling technologies are employed to create injection and production wells in the artificially created reservoir

Environmental and Economic Aspects of Geothermal Development

Environmental Considerations

- Geothermal energy is considered a renewable and low-carbon energy source, as the heat extracted from the Earth is continuously replenished by natural processes
- Geothermal power plants have a relatively small land footprint compared to other energy sources, and they can provide baseload power with high capacity factors

• Geothermal fluids can contain dissolved gases, such as carbon dioxide and hydrogen sulfide, which may need to be properly managed to minimize air pollution

• The extraction of geothermal fluids can lead to subsidence if not properly managed, as the removal of fluids can cause the ground to sink

• Induced seismicity is a potential concern in geothermal development, particularly in EGS projects, where hydraulic stimulation techniques are used to enhance reservoir permeability

Economic Factors

• The initial capital costs of geothermal projects can be high due to the need for exploration, drilling, and plant construction

• However, geothermal plants have low operational costs and long lifetimes (30-50 years), which can make them economically competitive in the long run

• Government incentives (tax credits, grants) and supportive policies can help to reduce the financial risks and encourage geothermal energy development

• The economic viability of geothermal projects depends on factors such as the resource quality, location (proximity to transmission lines), market conditions (electricity prices), and the availability of transmission infrastructure

• Geothermal energy can provide a stable and predictable source of revenue for local communities through job creation, tax revenues, and royalty payments

Unit 7 – Geophysical Well Logging

7.1 Principles of well logging and its applications

Geophysical Well Logging Principles

Well logging measures rock and fluid properties inside boreholes, giving you a continuous profile of what's happening underground. It's one of the most direct ways to characterize subsurface geology, estimate hydrocarbon reserves, and guide drilling decisions. The data you get from well logs feeds into nearly every stage of exploration and production.

Measurement and Recording of Physical Properties

A logging tool (called a sonde) is lowered into the borehole on a wireline cable or conveyed on drill pipe. As the tool moves through the hole, it records physical properties of the surrounding rock and fluids either continuously or at set depth intervals. The recorded data is transmitted to the surface and plotted as a log curve, with depth on the vertical axis and the measured property on the horizontal axis.

Each logging tool exploits a different physical interaction with the formation:

- Electrical conductivity (resistivity logs)
- Acoustic wave propagation (sonic logs)
- Nuclear reactions and radioactive decay (gamma ray, density, and neutron logs)
- Electromagnetic response (some specialized tools)

The result is a suite of curves that, taken together, let you reconstruct the geology around the borehole in considerable detail.

Objectives and Applications

The core goal of well logging is to determine lithology, porosity, permeability, and fluid content of subsurface formations. Beyond that, well logs serve several broader purposes:

- Stratigraphic correlation: Identifying and matching rock units, unconformities, and marker beds across multiple wells in a field or basin.
- Reservoir modeling: When combined with seismic data and core analysis, log data improves the accuracy of 3D reservoir models.
- Drilling optimization: Real-time logging (logging while drilling, or LWD) helps drillers make decisions about well trajectory and casing points as they go.

No single log type gives you the full picture. The real power comes from integrating multiple log types with other subsurface datasets.

Well Log Types and Applications

Lithology and Porosity Logs

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Gamma ray (GR) log: Measures the natural radioactivity of formations. Shales are typically high-GR because they contain radioactive potassium, uranium, and thorium, while clean sandstones and carbonates read low. This makes the GR log your go-to tool for distinguishing shale from reservoir rock and for correlating between wells.

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Density log: A radioactive source emits gamma rays into the formation, and detectors measure how many scatter back. The count rate is inversely related to the formation's bulk density (ρ_b). Since mineral grain density is roughly known, you can calculate porosity: $\phi = \frac{\rho_{ma} - \rho_b}{\rho_{ma} - \rho_f}$, where ρ_{ma} is matrix density and ρ_f is fluid density.

-

Neutron log: Bombards the formation with neutrons and measures how quickly they slow down. Hydrogen is the most effective neutron moderator, so the neutron log responds primarily to hydrogen content, which in most cases reflects porosity (since pore fluids contain hydrogen). Gas zones are an important exception because gas has lower hydrogen density than liquid, causing the neutron log to read anomalously low porosity.

-

Sonic log: Measures the travel time of compressional acoustic waves through the formation. Faster travel times correspond to denser, less porous rock. Sonic data also feeds into mechanical property calculations (like Young's modulus) used in wellbore stability and hydraulic fracture design.

Fluid and Borehole Condition Logs

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Resistivity logs: Measure the electrical resistivity of the formation. Water (especially saline formation water) conducts electricity well, so water-saturated rock has low resistivity. Hydrocarbons are electrical insulators, so hydrocarbon-bearing zones show higher resistivity. Different resistivity tools read at different depths of investigation (shallow, medium, deep), which helps distinguish invaded zones near the borehole from the undisturbed formation.

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Caliper log: Measures the actual diameter of the borehole. Where the hole is washed out or caved in, the caliper reads larger than bit size. This matters because borehole irregularities affect the accuracy of other logs, especially density and neutron measurements. Always check the caliper before trusting other log readings in a given interval.

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Image logs: Produce high-resolution pictures of the borehole wall using either resistivity contrasts or acoustic reflections. These reveal bedding planes, fractures, vugs, and sedimentary structures that standard logs can't resolve. They're particularly valuable in fractured reservoirs and for determining structural dip.

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Density-neutron combination: Plotting density and neutron porosity together on the same track is a standard technique. In clean, liquid-filled formations the two curves track each other. When they separate (crossover), it often indicates gas, because gas reduces the neutron reading while having a smaller effect on the density reading. This gas crossover effect is one of the most reliable quick-look hydrocarbon indicators.

Subsurface Formation Characterization

Lithology and Porosity Determination

Interpreting well logs means translating measured physical properties into geological information. No single log uniquely identifies a rock type, so you combine several:

1. Start with the gamma ray to separate shale from non-shale intervals.
2. Use density and neutron logs together to estimate porosity and flag gas zones.
3. Add the sonic log as a third porosity estimate and for lithology discrimination (carbonates, sandstones, and evaporites have distinct sonic velocities).
4. Cross-plot techniques (e.g., density vs. neutron, or sonic vs. density) help resolve ambiguous lithologies. On a density-neutron crossplot, for instance, limestone, dolomite, and sandstone each fall along different trend lines.

For complex formations with mixed mineralogy, multi-mineral analysis uses simultaneous equations from multiple log inputs to solve for the volume fractions of several minerals and fluids at once. This is common in carbonate reservoirs and unconventional plays where simple two-component models break down.

Stratigraphic Analysis and Correlation

Well logs are one of the primary tools for building a stratigraphic framework across a field or basin.

- Gamma ray logs are the most widely used for correlation because they respond to depositional environment. A fining-upward sequence (channel fill, for example) shows a characteristic bell-shaped GR pattern, while a coarsening-upward sequence (prograding shoreline) shows a funnel shape.
- Density and neutron logs help identify marker beds, such as a tight limestone or anhydrite layer, that produce distinctive spikes recognizable across wells.
- Unconformities often show up as abrupt shifts in log character, sometimes accompanied by a caliper anomaly if the contact is mechanically weak.

Correlating logs between wells builds a picture of how formations thicken, thin, pinch out, or change facies laterally. Combining this with seismic data and core descriptions gives you a much more robust understanding of the depositional history and structural setting.

Well Logging in Hydrocarbon Exploration

Reservoir Identification and Evaluation

Well logging provides the critical measurements needed to answer the basic exploration questions: Is there a reservoir? Does it contain hydrocarbons? How much?

- Resistivity logs are the primary hydrocarbon indicator. A zone with high resistivity in a porous interval strongly suggests hydrocarbons rather than saline water. Water saturation (S_w) is calculated using Archie's equation: $S_w = \left(\frac{a \cdot R_w}{\phi^m \cdot R_t}\right)^{1/n}$, where R_w is formation water resistivity, R_t is true formation resistivity, ϕ is porosity, and a, m, n are empirically determined constants.
- Porosity logs quantify storage capacity. Higher porosity generally means more room for hydrocarbons, though permeability (the ability of fluids to flow) also matters and is harder to measure directly from

logs.

• Net pay determination combines thickness, porosity, and saturation data to estimate how much of a formation actually contributes producible hydrocarbons, using cutoff values for each parameter.

Field Development and Production Optimization

Once a discovery is made, well logs continue to play a central role:

• Well placement: Log data from exploration and appraisal wells guides the positioning of development wells to target the best reservoir intervals.

• Completion design: Logs identify which zones to perforate and help engineers decide between open-hole and cased-hole completions.

• Fluid contacts: The oil-water contact (OWC) and gas-oil contact (GOC) can be picked from resistivity and porosity log responses, which is essential for volumetric reserve estimates.

• Time-lapse monitoring: Running logs in producing wells over time (e.g., pulsed neutron logs) tracks changes in fluid saturation, helping engineers monitor sweep efficiency and identify bypassed pay.

Integration of log data with seismic and core analysis remains the standard workflow for building and updating reservoir models. Advances in logging technology, including high-resolution imaging tools and logging-while-drilling systems, continue to improve characterization of complex and unconventional reservoirs such as shale gas and tight oil plays, where traditional log interpretation methods often need modification.

7.2 Acoustic and seismic logging

Acoustic and Seismic Well Logging

Acoustic and seismic logging uses sound waves to understand rock properties beneath the surface. These methods reveal porosity, fractures, and lithology, which are critical for resource exploration and reservoir management.

Wave Propagation Fundamentals

Sound waves traveling through rock behave differently depending on the material they encounter. P-waves (compressional) and S-waves (shear) each respond to density, elasticity, and fluid content in distinct ways. The velocity of these waves, how they attenuate, and how they reflect at boundaries all carry information about subsurface conditions.

P-wave velocity tends to be higher in dense, consolidated rock and lower in porous or fractured material. S-wave velocity is particularly sensitive to the rigidity of the rock matrix, since S-waves cannot propagate through fluids. The ratio between P-wave and S-wave velocities (V_p/V_s) is a key diagnostic for distinguishing rock types and saturation states.

At each interface between different rock layers, waves partially reflect and partially transmit. The acoustic impedance at a boundary (the product of rock density and wave velocity) determines how much energy reflects. Larger contrasts in impedance produce stronger reflections, which show up clearly in the logging data.

Logging Tool Design and Operation

Acoustic logging tools are lowered into the borehole on a wireline. A typical tool consists of a transmitter that generates a pulse of sound and one or more receivers spaced along the tool body that detect the arriving waveforms.

The transmitter fires a burst of acoustic energy into the formation. The sound radiates outward, interacts with the rock, and returns to the receivers after traveling through the near-borehole material. The travel time between the transmitted pulse and the received signal, along with the amplitude and waveform shape of the return, are all recorded.

Monopole tools use a single-point source and are standard for measuring compressional and shear slowness. Dipole tools add a directional source that generates flexural waves, which are especially useful for measuring shear velocity in slow formations where monopole shear arrivals may not be detectable. Array tools with multiple receivers improve the ability to separate different wave modes and reduce noise.

The spacing between transmitter and receivers matters. Wider spacing increases the depth of investigation into the formation but may reduce vertical resolution. Narrower spacing gives finer detail along the borehole wall but samples a smaller volume of rock.

Data Acquisition and Processing

Raw acoustic log data require processing to extract useful formation properties. The first step is waveform stacking, where multiple firings are combined to improve signal-to-noise ratio. Next, semblance analysis or slowness-time-coherence (STC) processing identifies the coherent arrivals corresponding to P-waves, S-waves, and Stoneley waves.

Slowness (the inverse of velocity, typically in μs/ft) is the primary measurement extracted from the data. Compressional slowness (DTc) and shear slowness (DTs) are computed from the identified arrivals. These values are then converted to velocities and used to derive mechanical and petrophysical properties.

Stoneley wave analysis provides additional information. These low-frequency waves propagate along the borehole interface and are sensitive to formation permeability. Their dispersion characteristics (how velocity changes with frequency) can be inverted to estimate near-borehole permeability.

Quality control is applied throughout. Bad data from tool eccentricity, borehole washouts, or cycle skipping are flagged and either corrected or excluded. Caliper logs (which measure borehole diameter) are used alongside the acoustic data to identify intervals where borehole conditions may have compromised the measurements.

Formation Evaluation from Acoustic Data

Acoustic data feed directly into formation evaluation. Porosity is estimated from compressional slowness using empirical relationships such as the Wyllie time-average equation, which relates transit time to the proportion of solid matrix versus pore fluid along the wave path.

Lithology identification uses the combination of DTc and DTs. Different minerals (quartz, calcite, dolomite, clay) have characteristic slowness values. Cross-plotting compressional versus shear slowness, or overlaying acoustic-derived porosity with neutron and density porosity, helps resolve ambiguous lithologies.

Mechanical properties are derived from the elastic wave velocities. Young's modulus, Poisson's ratio, bulk modulus, and shear modulus are all computed from V_p, V_s, and bulk density. These properties are essential for wellbore stability analysis, hydraulic fracture design, and sand production prediction.

Fracture detection relies on several indicators in the acoustic data. Fractured intervals often show anomalously low velocities, high attenuation, and scattered waveform energy. Shear wave splitting (where a shear wave divides into fast and slow components aligned with fracture orientation) is a direct indicator of aligned fractures or stress anisotropy. Comparing acoustic-derived fracture indicators with image log data provides a more complete fracture characterization.

Fluid substitution modeling uses the Gassner equations to predict how acoustic velocities would change if the pore fluid changed (for example, from brine to gas). This is valuable for evaluating different saturation scenarios and for calibrating seismic data with well log measurements.

Integration with Other Measurements

Acoustic logs are most powerful when combined with other logging measurements. Gamma ray logs identify shale content, which affects acoustic velocities and must be accounted for in porosity calculations. Density logs provide the bulk density needed to compute acoustic impedance and mechanical properties. Resistivity logs give independent fluid saturation information that complements the acoustic-derived properties.

Neutron-density-acoustic combinations are a standard approach for resolving porosity and lithology simultaneously. Each measurement responds differently to matrix composition and pore fluid, so combining them reduces ambiguity.

Vertical seismic profile (VSP) data tie the borehole acoustic measurements to surface seismic surveys. The acoustic log provides a detailed velocity profile along the well, which is used to build a synthetic seismogram that correlates well log features with seismic reflectors. This well-to-seismic tie is fundamental for extending borehole observations laterally across the field.

Checkshot surveys (where a surface seismic source fires into downhole receivers at selected depths) provide an independent time-depth relationship that validates the integrated acoustic velocity profile. Discrepancies between checkshot times and acoustic log transit times may indicate intervals of dispersion, near-borehole alteration, or measurement error that need correction.

7.3 Electrical and electromagnetic logging

Electrical and Electromagnetic Logging Principles

Electrical and electromagnetic logging methods measure the resistivity and conductivity of subsurface formations to determine lithology, fluid content, and saturation. These measurements are central to identifying hydrocarbon-bearing zones and estimating how much oil, gas, or water fills the pore space. Because hydrocarbons are far more resistive than formation water, resistivity logs give you one of the most direct indicators of pay zones.

Two broad approaches exist. Electrical logging applies a current directly to the formation and measures potential differences. Electromagnetic logging induces eddy currents in the formation and measures the resulting electromagnetic response. Both approaches are shaped by the same formation properties, but each has strengths in different borehole and formation conditions.

Fundamental Concepts and Measurement Techniques

Formation resistivity depends on several factors working together:

- Lithology — sandstones, shales, and carbonates each have characteristic baseline resistivities
- Porosity — more pore space means more room for conductive fluids, generally lowering resistivity

- Fluid type — water (especially saline water) conducts electricity well; oil and gas do not
- Saturation — the fraction of pore space filled by each fluid directly controls the bulk resistivity

The depth of investigation (how far into the formation the tool "sees") and vertical resolution (the thinnest bed the tool can distinguish) both depend on electrode or coil spacing, signal frequency, and formation resistivity. Wider spacing and lower frequencies probe deeper but resolve less detail. Tighter spacing and higher frequencies give sharper vertical resolution but read shallower.

Factors Influencing Formation Resistivity and Conductivity

The dominant control on formation resistivity is the type and volume of fluid in the pores. Saline formation water is highly conductive, so water-saturated rocks have low resistivity. Hydrocarbons replace conductive water with a resistive fluid, driving resistivity up.

Archie's Law quantifies this relationship for clean (clay-free), consolidated formations:

$$R_t = a\,R_w\,\phi^{-m}\,S_w^{-n}$$

where:

- R_t = true formation resistivity
- R_w = resistivity of formation water
- ϕ = porosity
- S_w = water saturation (fraction of pore space filled with water)
- a = tortuosity factor (often ~1)
- m = cementation exponent (typically 1.8–2.0 for sandstones)
- n = saturation exponent (typically ~2)

Rearranging for S_w is the most common use: if you know R_t from the log, R_w from water samples or SP logs, and ϕ from porosity logs, you can solve for water saturation. Whatever pore space isn't water is assumed to be hydrocarbons.

Beyond fluids, mineralogy matters. Conductive minerals like clays and pyrite lower the measured resistivity, which can mask hydrocarbons and cause you to overestimate S_w. Non-conductive matrix minerals (quartz, calcite) have the opposite effect. Temperature also plays a role: hotter formation water conducts better, so resistivity decreases with depth as temperature rises.

Logging Methods and Applications

Conventional and Focused Resistivity Logs

Normal and lateral logs are the oldest electrical logging tools. They use different electrode arrangements and spacings to sample different depths into the formation:

- Normal logs (e.g., 16-inch short normal, 64-inch long normal) place the current and potential electrodes on the same tool. The spacing between them sets the depth of investigation. The 16-inch reads the invaded zone; the 64-inch reads deeper.
- Lateral logs place the current electrode above the potential electrodes, producing a more focused measurement with better vertical resolution.

Micro-resistivity logs read very close to the borehole wall. Tools like the micro-spherically focused log (MSFL) and micro-cylindrically focused log (MCFL) have tiny electrode spacings that resolve thin beds (down to a few inches) and characterize the flushed zone created by mud filtrate invasion.

Laterologs (e.g., dual laterolog, array laterolog) use guard electrodes to force current into the formation in a focused beam. This minimizes the influence of the borehole fluid and invasion zone. Laterologs work best in low-resistivity formations drilled with water-based muds, where unfocused tools would be overwhelmed by the conductive borehole environment.

Induction and Dielectric Logging Techniques

Induction logs work on a completely different principle. A transmitter coil generates an alternating magnetic field that induces eddy currents in the formation. A receiver coil then measures the secondary electromagnetic field produced by those eddy currents, which is proportional to formation conductivity.

Induction tools are the go-to choice when: - Formation resistivity is high (roughly >50 ohm-m) - The well is drilled with oil-based or non-conductive mud (where laterologs can't establish a current path through the borehole fluid)

Higher transmitter frequencies and shorter coil spacings improve vertical resolution but reduce the depth of investigation.

Dielectric logs measure the dielectric permittivity of the formation rather than its resistivity. Water has a dielectric constant of about 80, while most hydrocarbons fall between 2 and 5, and rock matrix is around 4–8. This large contrast makes dielectric logs especially useful in situations where resistivity alone is ambiguous, such as low-resistivity pay zones, freshwater formations, or complex lithologies where Archie's Law breaks down.

Resistivity and Conductivity Logs for Fluid Analysis

Identification of Potential Pay Zones

A hydrocarbon-bearing zone shows up as a resistivity increase relative to surrounding water-wet formations. But resistivity alone doesn't tell you everything. Combining resistivity with other log types sharpens the interpretation:

- High resistivity + low neutron porosity + low bulk density → likely gas (gas has very low hydrogen index and low density)
- High resistivity + moderate neutron porosity + moderate density → likely oil
- Low resistivity throughout → likely water-bearing (or shaly formation requiring further analysis)

This multi-log approach is standard practice because no single measurement uniquely identifies fluid type.

Estimation of Fluid Saturation and Hydrocarbon Type

Once you've identified a potential pay zone, the next step is estimating S_w using Archie's Law. The hydrocarbon saturation is then $S_h = 1 - S_w$.

Invasion profiles provide additional diagnostic information. When drilling fluid invades the formation, it displaces some of the original pore fluid near the borehole. Comparing shallow-reading and deep-reading resistivity tools reveals a characteristic profile:

- Step profile (shallow resistivity < deep resistivity): Typical of a water-bearing zone where conductive mud filtrate has invaded, making the near-borehole zone even more conductive than the uninvaded formation.

- Ramp profile (shallow resistivity > deep resistivity): Typical of a hydrocarbon-bearing zone where mud filtrate (often more resistive than formation water) has displaced some hydrocarbons near the borehole, raising the shallow resistivity above the deep reading.

Recognizing these patterns helps confirm whether a resistivity anomaly reflects true formation fluids or just invasion effects. Dielectric logs can further resolve ambiguity, since the large dielectric contrast between water and hydrocarbons persists regardless of salinity.

Induction and Laterolog Techniques for Resistivity Evaluation

Selection of Appropriate Logging Techniques

Choosing between induction and laterolog tools comes down to three main factors:

1. Expected formation resistivity — Induction tools perform best in resistive formations (>50 ohm-m). Laterologs perform best in conductive formations (<50 ohm-m).

2. Drilling mud type — Induction tools work in oil-based or air-drilled holes where there's no conductive path through the borehole fluid. Laterologs require a conductive (water-based) mud to establish a current circuit.

3. Desired resolution and depth of investigation — Array tools in both families offer multiple spacings, giving you both shallow and deep readings simultaneously.

In practice, many logging programs run both tool types (or array versions that span a wide resistivity range) to get the most complete picture.

Environmental Corrections and Interpretation Challenges

Raw log readings must be corrected for environmental effects before you can extract true formation resistivity:

- Borehole size — Larger boreholes expose more mud to the tool, pulling the apparent resistivity toward the mud resistivity. Caliper data is used to apply the correction.

- Mud resistivity — Conductive muds decrease apparent resistivity readings. The mud resistivity (measured at surface and corrected to downhole temperature) is a required input.

- Temperature — Formation fluid resistivity drops as temperature increases. You need a reliable temperature profile to correct both the mud and formation water resistivity values.

Service companies provide correction charts (or software equivalents) for each tool. Skipping these corrections can introduce significant error, especially in large or rugose boreholes.

Common interpretation challenges include:

- Thin beds — Beds thinner than the tool's vertical resolution get averaged with surrounding layers, suppressing resistivity peaks and making pay zones look less promising than they are. High-resolution tools or deconvolution techniques help.

- Conductive minerals — Clays and pyrite conduct electricity through pathways that have nothing to do with pore fluids. Archie's Law assumes all conductivity comes from pore water, so shaly formations require modified equations (e.g., Waxman-Smits or dual-water models) to avoid overestimating S_w.

- Complex lithologies — Carbonates, fractured formations, and vuggy porosity systems don't follow the simple pore geometry Archie's Law assumes. These require integration of image logs, core data, and specialized interpretation workflows.

7.4 Nuclear logging methods

Nuclear logging methods measure formation properties by analyzing how radiation interacts with subsurface materials. These techniques provide data on lithology, porosity, and fluid content that are central to evaluating reservoir potential and guiding production decisions.

Nuclear Logging Methods

Types and Applications

Nuclear logging uses the interaction of radiation with matter to probe subsurface formations. The three main types each target different formation properties:

- Gamma ray logging measures the natural radioactivity of formations. It's primarily used for lithology identification and stratigraphic correlation.
- Neutron logging uses an artificial neutron source to measure hydrogen content, which relates directly to porosity and fluid type.
- Density logging uses an artificial gamma ray source to measure bulk density, from which porosity and lithology can be derived.

The choice of method depends on what formation property you need to characterize. In practice, these tools are almost always run together so their responses can be compared.

Principles and Interactions

Each type of nuclear log relies on a different radiation-matter interaction:

- Gamma rays interact with formation electrons through three mechanisms: photoelectric absorption, Compton scattering, and pair production. Which mechanism dominates depends on the gamma ray energy.
- Neutrons lose energy primarily through elastic collisions with hydrogen nuclei, since hydrogen has nearly the same mass as a neutron and therefore absorbs the most energy per collision.

The measured radiation response is influenced by:

- Formation lithology and mineralogy
- Porosity and the type of fluid filling the pore spaces
- Borehole conditions (diameter, fluid type, and casing)

Proper calibration and environmental corrections are essential for accurate interpretation. Without them, borehole effects can introduce significant errors into your readings.

Gamma Ray Logging for Lithology

Measurement Principles

Gamma ray logging detects the natural gamma radiation emitted by radioactive isotopes in the formation. The three primary contributors are potassium-40 (^{40}K), uranium (U), and thorium (Th).

The tool uses a scintillation detector that counts gamma rays over a specified time interval. Readings are reported in API units (American Petroleum Institute units), a standardized scale calibrated against a reference formation at the University of Houston.

The gamma ray response depends on the concentration of radioactive elements:

- Shales tend to produce high gamma ray readings because clay minerals concentrate potassium, uranium, and thorium.
- Clean sandstones and carbonates typically produce low readings because they contain fewer radioactive elements.

A spectral gamma ray tool can separate the contributions of K, U, and Th individually, which is useful when uranium enrichment (from organic matter or fractures) would otherwise make a clean formation look shaly.

Lithology Identification and Stratigraphic Correlation

The most common use of the gamma ray log is distinguishing shale from non-shale formations. Because shale content varies predictably with depositional environment, gamma ray logs are also valuable for:

- Depth matching other well logs to a common reference
- Identifying formation boundaries and marker beds
- Stratigraphic correlation between wells across a field

In a clastic sedimentary sequence, for example, gamma ray logs reveal sand-shale alternations as repeated swings between low and high API values. Abrupt shifts in the gamma ray baseline can indicate unconformities or sequence boundaries.

Neutron Logging for Porosity

Measurement Principles

Neutron logging uses an artificial source, typically americium-beryllium ($^{241}Am\text{-}Be$), which emits high-energy (fast) neutrons into the formation. These neutrons undergo elastic collisions with atomic nuclei and progressively lose energy. Hydrogen nuclei are by far the most effective at slowing neutrons because of the near-equal mass between a neutron and a proton.

Once slowed to thermal energies, neutrons are captured by formation nuclei, which then emit capture gamma rays. The tool measures either:

- The count rate of slowed-down (thermal or epithermal) neutrons (neutron-neutron logging)
- The count rate of capture gamma rays (neutron-gamma logging)

The neutron log response is governed by the hydrogen index (HI) of the formation, which quantifies the amount of hydrogen per unit volume relative to pure water. Higher hydrogen content means more neutron moderation near the source, so fewer neutrons or gamma rays reach the far detector. This inverse relationship is converted to a porosity reading.

Porosity and Fluid Content Determination

In clean, water-filled formations, the neutron log gives a reliable measure of porosity because nearly all the hydrogen resides in the pore fluid.

However, the neutron tool responds to all hydrogen, not just water. This creates important effects:

- Gas zones contain less hydrogen per unit volume than water, so the neutron log reads a porosity lower than the true porosity.
- Shale intervals contain bound water in clay minerals, which adds hydrogen and makes the neutron porosity read higher than the actual effective porosity.

Neutron logs are routinely combined with density logs to resolve these ambiguities. In a gas-bearing sandstone, for instance, the neutron log reads low while the density log reads high porosity, producing a characteristic crossover on the log display.

Density Logging for Formation Properties

Measurement Principles

Density logging uses a gamma ray source, typically cesium-137 (^{137}Cs), to emit medium-energy gamma rays into the formation. These gamma rays lose energy through Compton scattering off formation electrons.

The tool has two detectors at different spacings from the source. The count rate at each detector depends on the electron density of the formation, which is closely proportional to bulk density (ρ_b) for most common rock-forming minerals.

Bulk density reflects three components:

- Matrix density (ρ_{ma}): the density of the solid rock framework
- Porosity (ϕ): the fraction of pore space
- Fluid density (ρ_f): the density of whatever fills the pores

These are related by the bulk density equation:

$\rho_b = \phi \cdot \rho_f + (1 - \phi) \cdot \rho_{ma}$

Porosity and Lithology Determination

Rearranging the bulk density equation gives porosity directly:

$\phi = \frac{\rho_{ma} - \rho_b}{\rho_{ma} - \rho_f}$

You need to know (or assume) the matrix density and fluid density. Common matrix densities:

- Sandstone (quartz): ~2.65 g/cm³
- Limestone (calcite): ~2.71 g/cm³
- Dolomite: ~2.87 g/cm³

Lower bulk density values correspond to higher porosity. When gas is present, the fluid density drops below that of water (~1.0 g/cm³), so the density log reads a porosity higher than the neutron log. This is the opposite of what the neutron tool does in gas zones, which is exactly why the two logs are plotted together.

Density logs also help distinguish lithology. In a carbonate reservoir, for example, you can differentiate dense, low-porosity limestone from more porous dolomite intervals based on their characteristic density ranges.

Interpreting Nuclear Log Data

Formation Characterization

Nuclear logs together provide a detailed picture of the subsurface:

1. Gamma ray logs delineate shale from potential reservoir rock (sandstone or carbonate).
2. Neutron and density logs quantify porosity within those reservoir intervals.
3. Fluid type in the pore spaces is inferred from how neutron and density porosities compare.

Porosity is a critical parameter for assessing storage capacity. Without reliable porosity estimates, volumetric calculations of hydrocarbons in place are unreliable.

Hydrocarbon Identification

The separation between neutron and density porosity curves is the primary nuclear-log indicator of gas or light hydrocarbons. This is called the gas effect:

- In a water-filled zone, neutron and density porosities track each other closely.
- In a gas-filled zone, neutron porosity decreases (less hydrogen) while density porosity increases (lower bulk density), creating a visible crossover.

A zone that shows low gamma ray values (clean reservoir rock), reasonable porosity, and significant neutron-density crossover is a strong candidate for a hydrocarbon-bearing interval.

In a sandstone reservoir, a zone with low gamma ray values, high density porosity, and low neutron porosity is a classic gas signature.

Integration with Other Data

Nuclear logs should never be interpreted in isolation. Combine them with:

- Resistivity logs to confirm hydrocarbons (hydrocarbons increase resistivity relative to brine)
- Sonic logs to cross-check porosity estimates using an independent measurement
- Geological and seismic data to verify lateral continuity of the reservoir

A potential pay zone flagged by nuclear logs gains confidence when it also shows high resistivity and ties to a coherent seismic reflection. This multi-log, multi-data approach reduces the risk of false positives and supports better drilling and completion decisions.

Unit 8 – Remote Sensing and Geospatial Analysis

8.1 Principles of remote sensing and its applications in geophysics

Principles of Remote Sensing in Geophysics

Fundamentals of Remote Sensing

Remote sensing is the acquisition of information about an object or phenomenon without making physical contact, typically from aircraft or satellites. In geophysics, it provides a way to study large areas of Earth's surface quickly and repeatedly, which makes it indispensable for mapping, exploration, and monitoring.

All remote sensing relies on the interaction of electromagnetic radiation with matter. When EM radiation hits a surface, four things can happen: it can be reflected, absorbed, transmitted, or emitted. Different materials interact with radiation differently across the spectrum, and that's what allows us to distinguish them from a distance.

Remote sensing systems fall into two categories:

- Passive systems detect natural energy that is either reflected (like sunlight bouncing off rock) or emitted (like thermal radiation from warm ground). They depend on an external energy source, usually the Sun.
- Active systems generate their own energy, send it toward the target, and then measure what comes back. Radar and LiDAR are the main examples. Because they supply their own illumination, they can operate at night and, in the case of radar, through cloud cover.

Key Characteristics of Remote Sensing Systems

Three types of resolution define what a remote sensing system can do:

- Spectral resolution is the number and width of wavelength bands the sensor records. A sensor with many narrow bands (high spectral resolution) can distinguish materials that look similar in just a few broad bands. This matters when you're trying to tell one mineral from another.
- Spatial resolution is the size of the smallest feature the sensor can resolve, usually expressed as meters per pixel. A 1 m resolution image can pick out individual buildings; a 250 m resolution image cannot. Higher spatial resolution means more detail, but also larger data volumes and typically smaller coverage areas.
- Temporal resolution is how often the sensor revisits the same area. A satellite with a 16-day revisit cycle gives you a new image of a location roughly twice a month. Higher temporal resolution is critical for tracking dynamic processes like volcanic eruptions, flooding, or deforestation.

There's almost always a trade-off among these three. A sensor with very high spatial resolution often covers a smaller swath and may have fewer spectral bands. Choosing the right sensor for a geophysics project means deciding which resolution matters most for the question you're trying to answer.

Remote Sensing Techniques and Applications

Optical and Thermal Remote Sensing

Optical remote sensing works in the visible, near-infrared (NIR), and shortwave infrared (SWIR) portions of the spectrum, detecting sunlight reflected from Earth's surface.

- Multispectral systems record data in several discrete wavelength bands (typically 4–12). Each material has a characteristic spectral signature across these bands, so you can distinguish vegetation from bare soil from water. Landsat (with its ~30 m resolution and multiple bands) is a classic multispectral platform used heavily in geophysics.

- Hyperspectral systems collect data in hundreds of narrow, contiguous bands. This fine spectral sampling lets you identify specific minerals, map alteration zones, and characterize surface chemistry in ways that multispectral data cannot. For example, hyperspectral sensors can differentiate between clay minerals like kaolinite and montmorillonite based on subtle absorption features near 2.2 μm.

Thermal infrared (TIR) remote sensing measures energy emitted by the surface rather than reflected sunlight. Surface temperature and the thermal properties of materials (emissivity, thermal inertia) control the signal. TIR data are used in geothermal exploration to locate heat anomalies and in urban studies to map heat islands.

Active Remote Sensing Techniques

Radar remote sensing uses microwave-frequency pulses and records the backscattered return. The signal responds to surface roughness, geometry, moisture content, and dielectric properties.

- Synthetic Aperture Radar (SAR) exploits the motion of the platform to simulate a much larger antenna, producing high-resolution images. Because microwaves penetrate clouds and don't need sunlight, SAR works in any weather and at any time of day. Applications include monitoring oil spills, tracking sea ice, and detecting ships.

- Interferometric SAR (InSAR) compares the phase of two or more SAR images acquired at different times. Phase differences reveal surface displacement with millimeter-scale precision. This technique is widely used to measure ground deformation from earthquakes, track volcanic inflation, and monitor subsidence over aquifers or mining areas.

LiDAR (Light Detection and Ranging) fires rapid laser pulses and measures the round-trip travel time to calculate distance. The result is a dense 3D point cloud of the surface. LiDAR is especially valuable where you need bare-earth topography beneath vegetation, such as mapping fault scarps hidden under forest canopy, or characterizing forest structure.

Each technique has trade-offs. Radar penetrates clouds but has coarser spectral information. LiDAR gives extremely precise elevation data but covers smaller areas per flight. Optical sensors offer rich spectral detail but are blocked by clouds. Selecting the right tool depends on the geophysical question and the environmental conditions.

Interpreting Remote Sensing Data

Preprocessing and Enhancement

Raw remote sensing data contain distortions from the sensor, the platform's motion, and the atmosphere. Before analysis, you need to correct for these:

1. Radiometric correction removes sensor noise and atmospheric effects so that pixel values represent actual surface reflectance or emitted radiance rather than instrument artifacts.

2. Geometric correction (orthorectification) accounts for terrain relief, sensor viewing angle, and platform attitude so that the image aligns accurately with map coordinates.

Once the data are corrected, image enhancement techniques help bring out features of interest:

• Contrast stretching expands the range of pixel values to use the full display range, making subtle differences visible.

• Band ratios divide one spectral band by another to highlight specific materials. For example, the ratio of SWIR to NIR can emphasize clay mineral content while suppressing the effects of illumination variation.

• Histogram equalization redistributes pixel values to increase contrast in the most populated parts of the histogram.

Advanced Analysis Methods

• Spectral unmixing addresses the fact that a single pixel often contains more than one material. It decomposes the mixed signal into contributions from pure endmember spectra, estimating the fractional abundance of each material within the pixel. This is particularly useful for subpixel mapping of mineral abundances in heterogeneous terrain.

• Classification groups pixels with similar spectral characteristics into categories. In supervised classification, you provide training samples of known classes and the algorithm assigns all other pixels accordingly. In unsupervised classification, the algorithm clusters pixels statistically without prior labels, and you interpret the clusters afterward. Both approaches are used for land cover mapping and geological unit discrimination.

• Change detection compares images from different dates to identify where and how the surface has changed. Differencing, post-classification comparison, and time-series analysis are common approaches. Applications include tracking urban growth, glacier retreat, and post-disaster damage.

• Digital Elevation Models (DEMs) can be derived from stereo optical image pairs, InSAR, or LiDAR. DEMs provide the topographic base for deriving geomorphometric parameters like slope, aspect, and curvature, all of which feed into hazard assessment and geological interpretation.

Data Integration and Interpretation

Remote sensing data become most powerful when combined with other geospatial information: geological maps, gravity and magnetic surveys, borehole logs, and field observations. Integrating these layers in a GIS lets you cross-validate interpretations and build a more complete picture of subsurface structure and surface processes.

Sound interpretation also requires understanding the limitations. Atmospheric conditions, sensor calibration, mixed pixels, and temporal gaps all introduce uncertainty. Knowing what your data can and cannot tell you is just as important as knowing the analysis techniques.

Solving Geophysical Problems with Remote Sensing

Geological Applications

• Geological mapping: Different rock types and structural features produce distinct spectral, textural, and topographic signatures. Multispectral and hyperspectral imagery can delineate lithological units, while SAR and DEM data reveal faults, folds, and lineaments that may be subtle in optical imagery alone.

• Mineral exploration: Hyperspectral data are especially effective at mapping hydrothermal alteration zones, which often surround ore deposits. Identifying minerals like alunite, kaolinite, or iron oxides from orbit helps target field exploration and reduces costs by narrowing the search area before drilling.

Natural Hazard Assessment and Monitoring

• Volcanic monitoring: TIR sensors detect thermal anomalies at active vents, while InSAR tracks centimeter-scale inflation or deflation of the volcanic edifice. SAR can also map lava flows and track ash plumes regardless of weather.

• Landslide hazard assessment: High-resolution optical imagery and LiDAR-derived DEMs identify landslide scars and map susceptible slopes. InSAR can detect slow creep on unstable hillsides before a catastrophic failure occurs, giving early warning for at-risk communities.

• Earthquake and tsunami hazard assessment: InSAR measures interseismic strain accumulation along faults, helping to identify segments with elevated seismic hazard. After an earthquake, satellite imagery supports rapid damage assessment and guides emergency response.

Environmental and Resource Management

• Groundwater exploration: Optical and TIR data reveal surface indicators of shallow groundwater, including vegetation patterns (phreatophytes), soil moisture anomalies, and thermal contrasts. These clues help target drilling locations and map potential recharge zones.

• Glacier and ice sheet monitoring: Repeat optical imagery tracks changes in glacier extent, SAR measures ice velocity, and satellite laser altimetry (e.g., ICESat-2) quantifies thickness changes. Together, these datasets constrain mass balance estimates that are central to understanding sea-level rise.

• Geothermal exploration: TIR remote sensing maps surface temperature anomalies associated with subsurface heat sources. Combined with hyperspectral mapping of hydrothermal alteration minerals, this helps identify promising sites for geothermal energy development.

8.2 Satellite and airborne remote sensing techniques

Remote Sensing Platforms and Capabilities

Satellite and airborne remote sensing let us observe Earth's surface, atmosphere, and subsurface properties without direct contact. These platforms collect electromagnetic radiation reflected or emitted by the surface, and the choice of platform determines your spatial resolution, coverage area, and how often you can revisit the same location.

Satellite Platforms

Satellite platforms orbit Earth and collect data at regular intervals, providing global coverage and repeated observations over time. Major missions include Landsat (30 m resolution, since 1972), MODIS (250 m to 1 km, near-daily coverage), and Sentinel-2 (10-20 m, 5-day revisit).

Satellites are classified by orbit type, and each orbit suits different applications:

- Geostationary orbit (e.g., GOES): The satellite matches Earth's rotation, staying fixed above one point at ~35,786 km altitude. This allows continuous monitoring of weather systems and atmospheric dynamics over a specific region, but spatial resolution is coarse.
- Polar/sun-synchronous orbit (e.g., Landsat, NOAA satellites): The satellite passes near both poles, gradually covering the entire globe. Sun-synchronous orbits cross the equator at the same local solar time each pass, which keeps illumination conditions consistent between acquisitions.

The choice of platform depends on the tradeoff between spatial resolution, temporal resolution (revisit time), and area of coverage. Weather forecasting needs frequent revisits (geostationary), while detailed land cover mapping needs finer spatial resolution (polar-orbiting).

Airborne Platforms

Airborne platforms (aircraft and drones) operate at lower altitudes, which gives them higher spatial resolution and more flexibility in when and where data is collected.

- Aircraft can carry optical, thermal, and radar sensors, and they're well suited for targeted surveys over mineral exploration sites, agricultural fields, or urban areas.
- Drones (UAVs) achieve even finer resolution (centimeter-scale) and can be deployed rapidly for small-area surveys or emergency response.
- Airborne platforms are particularly useful for detailed mapping, infrastructure monitoring, and environmental assessment where satellite resolution isn't sufficient.

The tradeoff: airborne data covers smaller areas and costs more per unit area than satellite data, but the spatial detail and scheduling flexibility are far superior.

Multispectral vs. Hyperspectral Imaging

Both techniques measure reflected electromagnetic radiation across multiple wavelength bands, but they differ in spectral resolution, which determines how finely you can distinguish materials.

Multispectral Imaging

Multispectral sensors capture data in a small number of discrete, relatively broad spectral bands (typically 3-10) spanning visible, near-infrared (NIR), and shortwave infrared (SWIR) wavelengths. Landsat 8, for example, has 11 bands, while Sentinel-2 has 13.

Because there are fewer bands, multispectral data is faster to process and easier to work with. It's the workhorse for:

- Land cover classification (forest vs. urban vs. water)
- Vegetation health monitoring (using NIR and red bands)
- Regional-to-global geological mapping

The limitation is that broad bands can't distinguish materials with similar overall reflectance but subtle spectral differences.

Hyperspectral Imaging

Hyperspectral sensors collect data across hundreds of narrow, contiguous bands (often 100-200+), producing a near-continuous spectral signature for every pixel. A spectral signature is the unique pattern of reflectance and absorption across wavelengths that acts like a fingerprint for a material.

This high spectral resolution lets you detect subtle variations that multispectral sensors miss:

- Distinguishing between mineral species with similar color but different absorption features (e.g., kaolinite vs. montmorillonite clays)
- Identifying soil composition and contamination
- Assessing water quality parameters like chlorophyll concentration

Processing hyperspectral data is more computationally demanding. Common analysis techniques include:

- Principal Component Analysis (PCA): Reduces the hundreds of bands to a smaller number of uncorrelated components that capture most of the variance, making the data more manageable.
- Spectral unmixing: Decomposes each pixel's spectrum into contributions from known "endmember" spectra, estimating the fractional abundance of each material within the pixel.

Key distinction: Multispectral imaging tells you what general category a surface belongs to. Hyperspectral imaging can tell you what specific material is present.

Deriving Geophysical Parameters

Raw remote sensing data needs significant processing before it yields meaningful geophysical information. The workflow moves from pre-processing through parameter retrieval to validation.

Pre-processing and Corrections

Before any analysis, three types of corrections are applied:

1. Radiometric correction: Accounts for sensor calibration differences and removes instrument noise, ensuring measurements are consistent across sensors and time periods.
2. Atmospheric correction: Removes the effects of atmospheric scattering (Rayleigh, aerosol) and gas absorption (water vapor, ozone) to recover true surface reflectance. Without this step, two images of the same surface taken on different days could look very different.
3. Geometric correction (orthorectification): Corrects for terrain displacement, sensor viewing angle, and Earth's curvature so that pixels align accurately with real-world coordinates and other geospatial datasets.

Parameter Retrieval Algorithms

Once you have corrected surface reflectance or radiance, various algorithms extract geophysical parameters:

- Vegetation indices: The Normalized Difference Vegetation Index (NDVI) is calculated as:

$$NDVI = \frac{NIR - Red}{NIR + Red}$$

Healthy vegetation strongly reflects NIR and absorbs red light, so NDVI values near +1 indicate dense, healthy vegetation, while values near 0 indicate bare soil or water.

- Land surface temperature (LST): Derived from thermal infrared bands using split-window or single-channel algorithms that correct for surface emissivity and atmospheric effects.
- Soil moisture: Estimated from microwave (radar) data, since the dielectric constant of soil increases sharply with water content, changing the backscatter signal.
- Digital elevation models (DEMs): Generated from stereo optical imagery or radar interferometry (InSAR), representing surface topography at resolutions from ~30 m (SRTM) down to sub-meter (lidar).
- Spectral unmixing: Linear spectral unmixing estimates the fractional abundance of different materials within a mixed pixel by modeling the pixel's spectrum as a weighted sum of endmember spectra.

Time-Series Analysis and Validation

Repeated observations over time enable monitoring of dynamic processes:

- Detecting land cover change (deforestation, urban expansion)
- Tracking vegetation phenology (green-up, senescence cycles)
- Measuring surface deformation trends (subsidence, uplift)

Multitemporal data reveals trends, seasonal patterns, and anomalies that single-date imagery cannot capture.

Validation is essential. Derived parameters must be compared against independent reference data from field surveys, in-situ sensors, or ground-truthing campaigns. Common accuracy metrics include:

- Root mean square error (RMSE): Quantifies the average magnitude of prediction errors.
- Correlation coefficient (r or R^2): Measures how well remote sensing estimates track ground observations.

Remote Sensing Techniques for Geophysical Applications

Optical Remote Sensing

Optical sensors capture reflected sunlight in visible and infrared wavelengths. This is the most widely used remote sensing approach for surface characterization.

- Land cover classification uses algorithms like maximum likelihood, support vector machines, or random forests applied to multispectral imagery to map forests, urban areas, water bodies, and agricultural land.
- Vegetation monitoring relies on spectral indices (NDVI, EVI) to assess plant health, productivity, and seasonal cycles.
- Geological applications include mineral mapping using diagnostic absorption features, lithological discrimination between rock types, and structural analysis of faults and folds from lineament patterns.

The main limitation: optical sensors cannot see through clouds, and they only work during daylight.

Thermal Infrared Remote Sensing

Thermal sensors detect emitted radiation (typically 8-14 μm wavelength) rather than reflected sunlight, measuring surface temperature.

- Urban heat islands: Thermal imagery maps temperature differences between built-up areas and surrounding vegetation.
- Volcanic monitoring: Detects thermal anomalies at active volcanoes before and during eruptions.
- Evapotranspiration: LST is a key input for energy balance models that estimate water loss from agricultural fields.
- Thermal inertia mapping: Materials heat and cool at different rates. By comparing day and night thermal images, you can infer soil moisture and distinguish rock types based on their thermal properties.

Radar and Lidar Remote Sensing

These active sensors emit their own energy, which makes them independent of sunlight and (for radar) cloud cover.

Synthetic Aperture Radar (SAR): - Operates at microwave wavelengths that penetrate clouds, rain, and darkness. - InSAR (Interferometric SAR) compares the phase of radar returns from two passes over the same area to measure surface deformation with millimeter-scale precision. Applications include earthquake displacement mapping, volcanic inflation, and land subsidence from groundwater extraction. - Polarimetric SAR transmits and receives radar in different polarization states (HH, VV, HV), providing information on surface roughness, soil moisture, and vegetation biomass.

Lidar (Light Detection and Ranging): - Emits laser pulses and measures the return time to build high-resolution 3D point clouds of the surface. - Airborne Laser Scanning (ALS) produces DEMs at sub-meter resolution and can penetrate forest canopy to map bare-earth topography beneath. - Used for topographic mapping, forest inventory (canopy height, biomass estimation), fault scarp detection, and urban 3D modeling.

Multi-sensor Integration

No single sensor captures everything. Combining data from different platforms and sensor types provides complementary information and improves accuracy.

- Pan-sharpening fuses a high-resolution panchromatic band with lower-resolution multispectral bands to produce a sharp, color image.
- Optical + radar fusion improves land cover classification by combining spectral information (optical) with structural and moisture information (SAR).
- Lidar + hyperspectral integration enables simultaneous analysis of vegetation structure (3D canopy from lidar) and species composition (spectral signatures from hyperspectral).
- Decision-level fusion combines classification results from different sensors rather than merging raw data, letting each sensor contribute where it performs best.

These multi-sensor approaches leverage the strengths of each technique to build a more complete picture of geophysical processes than any single sensor could provide alone.

8.3 Geographic information systems (GIS) and their applications

GIS Fundamentals

Basic Concepts and Components

A Geographic Information System (GIS) is a computer-based system for capturing, storing, analyzing, and displaying geographically referenced data. It brings together hardware, software, databases, and trained personnel to enable spatial analysis and mapping. In geophysics, GIS serves as the central platform where you pull together diverse datasets (seismic surveys, gravity measurements, satellite imagery) and analyze them in a shared spatial framework.

Every GIS workflow moves through four main stages:

1. Data input — getting information into the system through digitizing, scanning, or GPS collection

2. Data storage and management — organizing data in databases (often geodatabases) so it can be efficiently queried and updated

3. Data manipulation and analysis — performing operations like overlay, buffering, and interpolation to extract meaning from the data

4. Data output — producing maps, reports, 3D visualizations, or interactive displays that communicate results

Data Organization and Models

GIS data is organized into layers, where each layer represents a specific theme or attribute (elevation, land use, geophysical properties, etc.). You stack and combine these layers to reveal spatial relationships. For example, overlaying a gravity anomaly layer on a geological map layer can highlight where subsurface structures correlate with known rock types.

Two primary data models are used:

- Vector (points, lines, and polygons) — best for discrete features like well locations (points), fault traces (lines), or survey boundaries (polygons)

- Raster (grid cells) — best for continuous phenomena like elevation, temperature, or magnetic field intensity. Each cell holds a single value, and the grid covers the entire area

Coordinate systems and map projections define the spatial reference framework for all data in a GIS. Because the Earth is roughly spherical but maps are flat, projections introduce distortions in area, shape, distance, or direction. Choosing the right projection matters: a poor choice can misalign datasets or distort spatial measurements. All layers in a GIS project need to share a common coordinate system, or at minimum be properly reprojected, so they line up correctly.

GIS for Geophysical Data

Data Management and Analysis

Geophysical projects generate large volumes of spatially referenced data: seismic surveys, well logs, gravity and magnetic measurements, electromagnetic soundings, and more. GIS provides the infrastructure to store this data in geodatabases, which enable efficient querying, updating, and sharing across research teams.

Spatial analysis techniques in GIS are especially valuable for estimating geophysical properties at locations where no measurements exist. Interpolation fills in the gaps between sampled points to create continuous surfaces. Kriging, a geostatistical interpolation method, is particularly common in geophysics because it not only estimates values but also quantifies the uncertainty of those estimates based on the spatial structure of the data.

GIS also enables overlay analysis, where you combine multiple geophysical datasets to look for correlations. A classic example: overlaying gravity anomaly data with magnetic anomaly data can help identify subsurface structures (like intrusions or basins) that produce signatures in both fields simultaneously.

Geostatistics and Visualization

Geostatistical tools built into GIS platforms help you assess the spatial dependence and variability of geophysical data before and after interpolation:

- Variograms model how data similarity changes with distance, which is essential for setting up kriging parameters correctly
- Spatial autocorrelation analysis quantifies whether nearby measurements are more similar than distant ones, aiding in data quality control and uncertainty assessment

GIS provides a wide range of visualization options for geophysical data:

- 2D maps — contour maps of gravity or magnetic anomalies, color-ramped surfaces of interpolated properties
- 3D models — subsurface block models integrating seismic reflection data with well log data
- Cross-sections — vertical slices through subsurface models showing layered structures
- Interactive displays — allowing users to rotate, zoom, and query features dynamically

These visualizations aren't just for presentation. They're active tools for data exploration, helping you spot patterns, anomalies, or errors that raw numbers alone won't reveal.

Remote Sensing Integration in GIS

Data Integration and Georeferencing

Remote sensing data (satellite imagery, airborne geophysical surveys, LiDAR) provides information about the Earth's surface and, in some cases, the shallow subsurface. GIS serves as the platform where you integrate this remote sensing data with ground-based measurements and other geospatial datasets.

Georeferencing is the process of assigning real-world coordinates to a remote sensing image so it aligns accurately with other GIS layers. This typically involves identifying ground control points (GCPs), which are locations visible in both the image and a reference dataset with known coordinates. Without proper georeferencing, overlaying remote sensing data on other layers produces misaligned, unreliable results.

Feature Extraction and Spectral Analysis

Once remote sensing data is georeferenced and loaded into GIS, you can extract geologically meaningful features from it:

- Lineaments (linear features that may indicate faults or fracture zones) can be digitized as vector lines and compared against geophysical anomaly maps

• Geomorphological patterns like drainage networks or landforms can be vectorized and analyzed for structural controls

Spectral analysis techniques within GIS enhance specific properties in multi-spectral remote sensing data:

• Band ratios divide one spectral band by another to highlight mineralogical differences (e.g., iron oxide vs. clay content)

• Principal component analysis (PCA) reduces redundant spectral information into a smaller number of components that capture the most variance, making subtle features more visible

Data fusion techniques like pan-sharpening combine high-resolution panchromatic (single-band) imagery with lower-resolution multi-spectral data. The result is an image with both high spatial resolution and rich spectral information, useful for detailed geological mapping.

GIS Mapping and Modeling for Geophysics

Thematic Mapping and Data Integration

GIS enables the creation of publication-quality, interactive maps that communicate geophysical data and interpretations clearly. Thematic mapping techniques include:

• Color ramping — assigning a gradient of colors to a range of values (e.g., blue-to-red for low-to-high Bouguer gravity anomaly)

• Data classification — grouping continuous values into discrete classes to highlight patterns or anomaly thresholds

A major strength of GIS is the ability to integrate geophysical data with contextual information like geological maps, well locations, roads, and infrastructure. This provides a comprehensive spatial context that supports more informed interpretation and planning.

3D Modeling and Decision Support

3D modeling capabilities in GIS allow you to build subsurface models by integrating multiple data types:

1. Start with surface geophysical data (gravity, magnetics, seismic reflection profiles)

2. Incorporate borehole data (lithology logs, geophysical well logs) as vertical constraints

3. Interpolate between data points to generate 3D volumes representing subsurface structures, reservoir properties, or fluid flow pathways

Multi-criteria decision analysis (MCDA) is a GIS-based approach that supports complex decision-making. It works by integrating and weighting various factors (geophysical anomaly strength, geological favorability, environmental constraints, infrastructure access) to produce a composite suitability map. This is commonly used for site selection in mineral exploration, geothermal development, or survey planning.

Web-based GIS platforms and cloud computing have expanded collaboration in geophysical projects. Teams can share maps, models, and raw data through online portals, making it possible for researchers in different locations to access and contribute to the same spatial datasets in near real-time.

8.4 Digital elevation models and terrain analysis

Digital Elevation Models: Principles and Methods

Digital elevation models (DEMs) are grid-based representations of Earth's surface where each cell stores an elevation value, producing a continuous picture of terrain across a landscape. They're foundational to geophysics because they let you quantitatively analyze slopes, drainage, landforms, and even subsurface structures when combined with other datasets.

DEMs are built from a range of data sources, and the method you choose determines the resolution, coverage, and accuracy you'll get.

Generation of Digital Elevation Models

Ground surveys use traditional instruments like total stations or GPS receivers to collect elevation points. These produce high-resolution DEMs but only for relatively small areas, since someone has to physically visit each measurement location.

Aerial photogrammetry takes overlapping aerial photographs and extracts elevation through stereoscopic analysis. Two photos of the same area from slightly different angles let you calculate depth, much like how your two eyes perceive depth. This scales to larger areas than ground surveys.

LiDAR (Light Detection and Ranging) fires laser pulses toward the ground and measures the round-trip travel time to calculate elevation. Because light travels at a known speed, the return time gives a precise distance. LiDAR produces very high-resolution DEMs and can even penetrate vegetation canopy to map the bare-earth surface beneath.

InSAR (Interferometric Synthetic Aperture Radar) compares the phase differences between two or more SAR images acquired from slightly different positions or at different times. Those phase shifts encode elevation information, making InSAR especially useful for detecting surface elevation changes over time.

SRTM (Shuttle Radar Topography Mission) used a radar system aboard the Space Shuttle in 2000 to collect near-global elevation data. The resulting DEM covers latitudes from 60°N to 56°S at approximately 30-meter resolution (1 arc-second) for the most recent release, making it one of the most widely used global elevation datasets.

Accuracy and Uncertainties in Digital Elevation Models

DEM accuracy depends on three main factors: the acquisition method, the spatial resolution of the grid, and the post-processing applied to the raw data. Higher-resolution DEMs capture finer terrain detail but demand more storage and computational power.

Common sources of error include:

- Vertical and horizontal accuracy limitations inherent to the sensor or survey method
- Data gaps from shadowed areas, dense vegetation, or sensor geometry
- Artifacts such as striping, pits, or spikes introduced during processing

Quantifying these uncertainties matters. If you're using a DEM with ±10 m vertical accuracy to map subtle slope changes of a few degrees, your results may not be reliable. Always check the reported accuracy metrics (typically RMSE, or root mean square error) before applying a DEM to a geophysical problem.

Terrain Analysis for Geomorphological and Hydrological Insights

Terrain analysis extracts quantitative information about surface shape and water flow from DEMs. These derived attributes connect topography to physical processes like erosion, runoff, and sediment transport.

Quantitative Characterization of Topographic Attributes

Slope measures the steepness of the terrain at each grid cell, typically expressed in degrees or percent. Aspect indicates the compass direction that the steepest slope faces. Together, slope and aspect control surface processes: steep south-facing slopes in the Northern Hemisphere receive more solar radiation, affecting weathering rates and vegetation patterns.

Curvature describes how the surface bends, and it comes in two flavors:

- Profile curvature is the rate of change of slope along the direction of steepest descent. Positive profile curvature indicates a concave (decelerating) slope; negative indicates convex (accelerating).
- Plan curvature is the rate of change of aspect perpendicular to the slope direction. It tells you whether water flow converges (hollows) or diverges (ridges).

Curvature analysis helps distinguish landform types. Convex surfaces often mark ridgelines and hilltops, concave surfaces indicate valleys and hollows, and planar surfaces suggest uniform hillslopes.

Hydrological Analysis and Derived Terrain Attributes

Hydrological analysis uses DEMs to trace where water goes. The standard workflow follows these steps:

1. Fill sinks in the DEM to remove small depressions that would trap flow unrealistically.
2. Calculate flow direction using an algorithm like D8, which assigns flow from each cell to whichever of its eight neighbors has the steepest downhill gradient.
3. Calculate flow accumulation by counting how many upslope cells drain through each cell. High accumulation values trace out stream channels.
4. Delineate catchments by identifying the contributing area that drains to a specific outlet point. Every cell upslope of that outlet, as determined by flow direction, belongs to the catchment.

From these basics, you can derive more advanced terrain attributes:

•

Topographic Wetness Index (TWI) combines slope and upstream contributing area: $TWI = \ln\left(\frac{a}{\tan\beta}\right)$ where a is the specific upslope contributing area (area per unit contour length) and β is the local slope angle. High TWI values flag areas prone to soil saturation, making it useful for mapping potential wetlands, recharge zones, and flood-susceptible areas.

•

Geomorphometric classification algorithms analyze local geometry to automatically identify landforms like ridges, valleys, peaks, and depressions. These rely on combinations of slope, curvature, and relative position within the landscape.

DEM Integration for Geophysical Interpretation

A DEM on its own shows you surface shape. Combining it with other datasets is where the real analytical power emerges.

Enhancing Understanding through Data Integration

Geological maps and structural data overlaid on DEMs reveal how subsurface geology shapes the landscape. Faults often produce linear scarps or offset drainage patterns visible in the topography. Folds can create systematic asymmetries in ridge profiles. Draping structural data onto a 3D DEM visualization makes these relationships much easier to interpret.

Remote sensing imagery (multispectral, hyperspectral) combined with DEMs enables mapping of surface materials, vegetation, and land use in a topographic context. For example, you can distinguish erosion-prone bare soil on steep slopes from vegetated stable areas, or track how land cover change correlates with terrain position.

Integrating Geophysical and Deformation Data

Gravity, magnetic, and seismic data gain context when viewed alongside surface topography. A gravity anomaly that aligns with a topographic depression might indicate a sedimentary basin, while a magnetic anomaly along a ridge could reflect an exposed igneous intrusion. The DEM helps constrain where subsurface features connect to surface expressions.

Surface deformation monitoring using GPS and InSAR data integrated with DEMs allows you to track ground movements from earthquakes, volcanic inflation, and landslides. The DEM provides the baseline topography against which deformation is measured and helps interpret whether observed movements correlate with specific terrain features like fault scarps or steep slopes.

Hydrological modeling benefits from combining DEMs with soil property maps, land cover data, and climate variables (precipitation, evapotranspiration). This integration enables simulation of surface runoff, subsurface flow, soil moisture dynamics, and flood hazard assessment.

When integrating datasets, pay attention to:

- Resolution mismatches between datasets (a 90 m DEM paired with 10 m land cover data requires resampling decisions)
- Coordinate system differences that need reprojection
- Error propagation, since uncertainties in each dataset compound in the integrated analysis

DEM Applications in Landslide Susceptibility and Groundwater Exploration

Landslide Susceptibility Mapping

Landslide susceptibility mapping estimates the spatial likelihood of landslide occurrence by combining topographic, geological, and environmental factors. DEMs are central to this because they provide the topographic predictors that most strongly control slope failure.

Key DEM-derived inputs for landslide models include:

- Slope gradient: Steeper slopes experience greater gravitational shear stress. Most landslide inventories show a strong correlation between failure locations and slopes above a critical threshold (often 25°–35°, depending on material).
- Aspect: Controls moisture and vegetation patterns that affect slope stability.
- Curvature: Concave slopes concentrate water, increasing pore pressure and failure risk.

- TWI and stream power index (SPI): Indicate where soil saturation and erosive energy are highest, both of which are landslide triggering factors. SPI is calculated as $SPI = a \times \tan\beta$, where a is the specific catchment area and β is slope.

Combining these topographic attributes with geotechnical data (soil shear strength, cohesion, groundwater table depth) and rainfall patterns produces susceptibility models that account for both surface and subsurface controls on stability. Statistical approaches like logistic regression or machine learning classifiers are commonly used to weight these factors against known landslide inventories.

Groundwater Exploration

DEMs help identify where groundwater accumulates, recharges, and discharges by revealing the topographic controls on subsurface water flow.

Drainage network delineation and catchment analysis highlight recharge zones (typically topographic highs with permeable soils) and discharge zones (topographic lows, springs, and gaining streams). Flow accumulation patterns point to areas where surface water concentrates and potentially infiltrates.

Integrating DEMs with geological and hydrogeological data strengthens the analysis:

- Lithology maps identify permeable vs. impermeable formations
- Fracture network data reveal preferential flow paths through bedrock
- Aquifer property data constrain where productive wells are most likely

The TWI, introduced earlier, also serves as a proxy for groundwater potential. Areas with high TWI values tend to have shallow water tables and higher soil moisture, making them priority targets for exploration.

Combining DEMs with geophysical survey data (electrical resistivity tomography, seismic refraction) further refines the picture. Resistivity data can delineate aquifer boundaries and estimate saturated thickness, while the DEM provides the surface topographic framework that controls recharge distribution. Together, these datasets guide well placement decisions and aquifer characterization far more effectively than any single data source alone.

Unit 9 – Geophysical Data Processing and Interpretation

9.1 Digital signal processing techniques

Digital Signal Processing for Geophysical Data

Digital signal processing (DSP) is how geophysicists turn raw field recordings into interpretable data. Geophysical measurements are always contaminated by noise, whether from instruments, environmental sources, or the Earth itself. DSP provides the mathematical tools to separate signal from noise, resolve subsurface features, and extract quantitative information from seismic, gravity, magnetic, and electromagnetic datasets.

This section covers the core DSP techniques you'll need: sampling theory, digital filtering, and windowing/tapering methods.

Overview of DSP in Geophysics

Most geophysical instruments record analog signals that are then digitized for computer processing. Once in digital form, you can apply a range of techniques to clean up and analyze the data.

The most common DSP techniques in geophysics include:

- Filtering to remove noise or isolate frequency bands of interest
- Convolution and correlation to compare signals or model system responses
- Fourier analysis to decompose signals into their frequency components
- Wavelet analysis to examine how frequency content changes over time

Which technique you choose depends on the data type, the noise characteristics, and what you're trying to extract. A seismic reflection survey calls for different processing than a magnetotelluric time series, for example. In practice, most processing workflows are implemented in MATLAB or Python using libraries like ObsPy or SciPy.

Sampling, Aliasing, and Nyquist Frequency

The Sampling Process

Sampling converts a continuous analog signal into a discrete digital signal by measuring the signal's amplitude at regular time intervals. The sampling rate (or sampling frequency), measured in hertz (Hz), is the number of samples taken per second.

The critical concept here is the Nyquist frequency, which equals half the sampling rate:

$$f_{Nyquist} = \frac{f_s}{2}$$

where f_s is the sampling rate. The Nyquist frequency is the highest frequency that can be accurately represented in the digital signal.

The Nyquist-Shannon sampling theorem states that to faithfully capture a signal, you must sample at a rate at least twice the highest frequency component present in that signal:

$f_s \geq 2f_{max}$

For example, if a seismic signal contains frequencies up to 125 Hz, you need a sampling rate of at least 250 Hz (a 4 ms sample interval).

Oversampling (sampling faster than required) improves signal quality and gives you a safety margin, but it increases data volume and processing time. In practice, geophysicists typically sample at 3 to 4 times the maximum frequency of interest.

Aliasing and Its Effects

Aliasing occurs when the sampling rate is too low to capture the highest frequency components in the signal. The result is that high-frequency energy gets "folded back" into lower frequencies, creating false spectral content that's indistinguishable from real data.

This is a serious problem because aliased signals can't be removed after digitization. Once the data is sampled, the damage is done.

To prevent aliasing:

1. Determine the highest frequency present in your analog signal
2. Set the sampling rate to at least twice that frequency (and preferably higher)
3. Apply an anti-aliasing filter (an analog low-pass filter) before digitization to remove any frequency content above the Nyquist frequency

The anti-aliasing filter is applied in the analog domain, before the signal reaches the analog-to-digital converter. This is the only reliable way to prevent aliasing, since you can't fix it in post-processing.

Digital Filters for Signal Enhancement

Types of Digital Filters

Digital filters selectively modify the frequency content of a signal. There are four basic filter types, each defined by which frequencies they pass or reject:

- Low-pass filter: Passes frequencies below a cutoff and attenuates higher frequencies. Use this to remove high-frequency noise (e.g., instrument chatter in a gravity survey).
- High-pass filter: Passes frequencies above a cutoff and attenuates lower frequencies. Useful for removing slow drift or regional trends from data.
- Band-pass filter: Passes only a specified frequency range and attenuates everything outside it. Common in seismic processing where you want to isolate the reflection signal band (say, 10–80 Hz).
- Notch filter: Rejects a very narrow frequency band. The classic application is removing power line interference at 50 Hz or 60 Hz.

When designing any filter, you need to specify:

1. The filter type (low-pass, high-pass, band-pass, or notch)
2. The cutoff frequency (or frequencies, for band-pass and notch filters)
3. The filter order, which controls how sharply the filter transitions between passband and stopband

Higher-order filters give sharper cutoffs but can introduce other issues like ringing or phase distortion.

FIR vs. IIR Filters

Digital filters fall into two broad categories based on their mathematical structure:

Finite Impulse Response (FIR) filters: - Their output depends only on current and past input values - Always stable, which makes them reliable - Can be designed with linear phase, meaning they don't distort the shape of waveforms (just delay them uniformly). This is important in seismic processing where waveform shape carries information. - The tradeoff: achieving a sharp cutoff requires many filter coefficients, which increases computation

Infinite Impulse Response (IIR) filters: - Their output depends on both input values and previous output values (feedback) - Achieve sharp frequency cutoffs with far fewer coefficients than FIR filters - Can be unstable if poorly designed, and they introduce nonlinear phase distortion, which alters waveform shapes - Common IIR designs include Butterworth (maximally flat passband), Chebyshev (sharper cutoff with passband ripple), and Bessel (best phase response)

Choose FIR filters when waveform fidelity and phase preservation matter (e.g., seismic reflection data). Choose IIR filters when computational efficiency is the priority and some phase distortion is acceptable.

You can evaluate filter performance by examining the frequency response (what frequencies pass or get attenuated), the impulse response (how the filter reacts to a spike input), and the phase response (how much phase shift the filter introduces at each frequency).

Windowing and Tapering Effects on Data

Windowing Techniques

In practice, you never process an infinitely long signal. You always select a finite segment of data for analysis. Windowing is the process of multiplying your data by a window function that defines which portion of the signal you're analyzing.

The simplest window is the rectangular window, which just chops out a segment with sharp edges. The problem is that these abrupt edges create artificial discontinuities, which produce spectral leakage: spurious frequency content that spreads energy from true spectral peaks into neighboring frequencies.

To reduce spectral leakage, you can use tapered window functions that smoothly roll off toward zero at the edges:

- Hamming window: Good reduction of the nearest sidelobe; widely used as a general-purpose window
- Hanning (Hann) window: Similar to Hamming but with slightly different sidelobe behavior; goes exactly to zero at the edges
- Blackman window: Stronger sidelobe suppression than Hamming or Hanning, but at the cost of a wider main lobe (reduced frequency resolution)

There's a fundamental tradeoff here: windows that suppress spectral leakage more aggressively also broaden spectral peaks, reducing your ability to distinguish closely spaced frequencies. You're always balancing spectral resolution against spectral leakage.

You can examine how windowing affects your data using the short-time Fourier transform (STFT), which applies the Fourier transform to successive windowed segments, or the continuous wavelet transform (CWT), which provides variable time-frequency resolution.

Tapering and Overlapping Windows

Tapering is the gradual reduction of signal amplitude at the edges of a data window. Common taper shapes include cosine tapers (also called Tukey windows) and Gaussian tapers. Tapering serves the same purpose as using a non-rectangular window function: it minimizes edge discontinuities and the resulting spectral leakage.

When computing spectral estimates from long time series, it's common to divide the data into overlapping segments. The Welch method is a standard approach:

1. Divide the time series into segments of a chosen length
2. Apply a window function to each segment
3. Compute the power spectrum of each windowed segment
4. Average the spectra across all segments

Using overlapping windows (typically 50% overlap) means that data near the tapered edges of one window falls near the center of the adjacent window, so no part of the signal is underweighted. This reduces the variance of the spectral estimate, giving you a smoother, more reliable spectrum.

The tradeoff with overlap percentage and window length is straightforward: longer windows give better frequency resolution, shorter windows give better time localization, and more overlap gives smoother estimates at the cost of increased computation.

9.2 Fourier analysis and filtering

Fourier Analysis in Geophysics

Principles and Applications

Fourier analysis works on a simple but powerful idea: any complex signal can be broken down into a sum of sinusoidal (sine and cosine) waves, each with its own frequency, amplitude, and phase. In geophysics, the signals you're working with are rarely simple. A seismic trace, for instance, contains reflections from multiple layers, noise from the surface, and instrument artifacts all stacked on top of each other. Fourier analysis lets you pull those apart by moving from the time domain (amplitude vs. time) into the frequency domain (amplitude vs. frequency).

The Fourier transform is the mathematical operation that performs this conversion. It comes as a pair:

- The forward Fourier transform converts from time to frequency domain
- The inverse Fourier transform converts back from frequency to time domain

This back-and-forth capability is what makes Fourier analysis so useful. You can transform your data into the frequency domain, manipulate specific frequency components (remove noise, isolate a signal band), and then transform back to get a cleaner time-domain signal.

Fourier analysis applies across nearly every branch of geophysical data processing:

- Signal processing of seismic, gravity, magnetic, and electromagnetic data
- Noise reduction by identifying and removing unwanted frequency components
- Spectral analysis to characterize the frequency content of a dataset

- Data compression by retaining only the most significant frequency components

Mathematical Foundations

The continuous forward Fourier transform converts a time-domain function $f(t)$ into its frequency-domain representation $F(\omega)$:

$$F(\omega) = \int_{-\infty}^{\infty} f(t)\, e^{-i\omega t}\, dt$$

The inverse Fourier transform recovers the original signal:

$$f(t) = \frac{1}{2\pi}\int_{-\infty}^{\infty} F(\omega)\, e^{i\omega t}\, d\omega$$

Here, ω is angular frequency (in radians per second), i is the imaginary unit, and the complex exponential $e^{-i\omega t}$ encodes both sine and cosine components via Euler's formula.

Several properties make the Fourier transform especially practical:

- Linearity: The transform of a sum of signals equals the sum of their individual transforms.
- Scaling: Compressing a signal in time spreads it out in frequency, and vice versa.
- Convolution theorem: Convolution in the time domain becomes simple multiplication in the frequency domain. This is why filtering is so much more efficient in the frequency domain. Instead of performing a computationally expensive convolution, you just multiply two spectra together.

Time vs. Frequency Domain Conversion

Discrete Fourier Transform (DFT)

Geophysical data is recorded as discrete samples, not continuous functions. The discrete Fourier transform (DFT) handles this by operating on a finite set of N samples.

The forward DFT is:

$$X[k] = \sum_{n=0}^{N-1} x[n]\, e^{-i\frac{2\pi}{N}kn}$$

The inverse DFT is:

$$x[n] = \frac{1}{N}\sum_{k=0}^{N-1} X[k]\, e^{i\frac{2\pi}{N}kn}$$

where $x[n]$ is the n-th time-domain sample and $X[k]$ is the k-th frequency-domain coefficient.

Computing the DFT directly requires $O(N^2)$ operations, which becomes prohibitively slow for large datasets. The fast Fourier transform (FFT) is an algorithm that exploits symmetries in the DFT to reduce this to $O(N\log N)$. For a seismic survey with millions of samples, this difference is the reason frequency-domain processing is feasible at all. Most FFT implementations work most efficiently when N is a power of 2, so data is often zero-padded to the next power of 2 before transformation.

Sampling and Aliasing

Before any Fourier analysis can happen, the continuous geophysical signal must be properly sampled and digitized. The Nyquist-Shannon sampling theorem sets the fundamental rule: the sampling frequency f_s must be at least twice the highest frequency component f_{max} present in the signal.

$f_s \geq 2f_{max}$

The frequency $f_s/2$ is called the Nyquist frequency. If the signal contains energy above this frequency, those components get "folded back" into lower frequencies during digitization. This is aliasing, and it's irreversible once the data is recorded. A 120 Hz signal sampled at 200 Hz, for example, would appear as an 80 Hz signal in the digitized data.

To prevent aliasing, anti-aliasing filters (analog low-pass filters) are applied before digitization to remove frequency content above the Nyquist frequency.

Power and Phase Spectra

The Fourier transform $F(\omega)$ is complex-valued, meaning it contains both magnitude and phase information. These are typically analyzed separately:

- The power spectrum is the squared magnitude $|F(\omega)|^2$. It shows how energy is distributed across frequencies. A seismic trace dominated by low-frequency reflections will show most of its power at low frequencies, while high-frequency noise will appear as elevated power at the high end.
- The phase spectrum is the argument (angle) of $F(\omega)$. It tells you the relative timing of each frequency component. Phase information is critical for preserving the shape and timing of waveforms during processing.

Together, the power and phase spectra fully characterize a signal's frequency content. You can reconstruct the original signal perfectly from these two spectra using the inverse Fourier transform.

Frequency-Domain Filtering

Types of Filters

Filtering in the frequency domain means selectively modifying specific frequency components of a signal. The four main filter types are:

-

Low-pass filters preserve frequencies below a cutoff and attenuate those above it. These are used to remove high-frequency noise or smooth data. In seismic processing, a low-pass filter might remove random noise above 80 Hz while keeping the reflection signal below that.

-

High-pass filters preserve frequencies above a cutoff and attenuate those below it. These remove low-frequency trends such as DC offsets or long-period drift. A high-pass filter at 5 Hz, for example, could remove slow instrument drift from a seismic record.

-

Band-pass filters preserve frequencies within a specified range and attenuate everything outside it. These are the most commonly used filters in seismic processing. A typical band-pass might keep frequencies between 10 and 60 Hz, isolating the useful reflection bandwidth.

-

Notch filters attenuate a narrow frequency band while preserving everything else. The classic application is removing 50 or 60 Hz power line interference from electromagnetic or seismic data.

Filter Design and Implementation

Designing a frequency-domain filter involves several choices:

1. Define the filter response by specifying cutoff frequencies, the width of the transition band (how sharply the filter rolls off), and the required stopband attenuation.

2. Choose a filter type. Common options include Butterworth (maximally flat passband), Chebyshev (sharper rolloff but with passband ripple), and elliptic filters (sharpest rolloff but with ripple in both passband and stopband).

3. Apply the filter. In practice, you multiply the signal's Fourier transform $F(\omega)$ by the filter's frequency response $H(\omega)$ to get the filtered spectrum $G(\omega) = F(\omega) \cdot H(\omega)$.

4. Transform back. Apply the inverse Fourier transform to $G(\omega)$ to obtain the filtered time-domain signal.

This multiplication-in-frequency approach is a direct application of the convolution theorem: multiplying spectra in the frequency domain is equivalent to convolving the signal with the filter's impulse response in the time domain, but computationally much faster.

Effects of Filtering on Geophysical Data

Signal-to-Noise Ratio Improvement

The primary goal of filtering is to improve the signal-to-noise ratio (SNR) by suppressing unwanted components:

- Low-pass filtering removes high-frequency random noise and measurement errors
- High-pass filtering removes low-frequency trends, DC offsets, or long-period drift
- Band-pass filtering isolates the frequency range where the desired signal is strongest, attenuating noise on both sides
- Notch filtering targets specific interference sources like power line noise or ground roll in seismic data

Potential Artifacts and Distortions

Every filter involves trade-offs. Applying a filter always changes the data, and not always in desirable ways:

- Low-pass filtering can blur sharp features and edges, reducing spatial or temporal resolution. Thin-bed reflections in seismic data, for instance, may be smeared together.
- High-pass filtering can amplify high-frequency noise, making the data appear grainy or noisy.
- Band-pass filtering can introduce Gibbs phenomenon (ringing artifacts) near sharp transitions or discontinuities, especially with steep filter rolloffs.
- Notch filtering can remove desired signal energy if the notch is too wide or poorly centered on the interference frequency.

Interpretation Considerations

Filtered data should always be compared against the original unfiltered data so you can assess what the filter actually did. A few practical guidelines:

- Choose filter parameters based on the known frequency content of both the signal and the noise. Examining the power spectrum before filtering helps you make informed choices.
- Consider the effects on amplitude, phase, and waveform shape. Zero-phase filters preserve event timing but aren't always achievable in practice; minimum-phase filters are causal but introduce phase distortion.
- Avoid over-filtering. Applying too many filters or using overly aggressive parameters can remove real geological signal along with the noise.
- Document and communicate your filter choices. Anyone interpreting the processed data needs to know what filters were applied, with what parameters, and what artifacts might have been introduced.

9.3 Inversion and modeling techniques

Forward vs Inverse Modeling in Geophysics

Defining Forward and Inverse Modeling

Forward and inverse modeling are two complementary ways of relating subsurface properties to geophysical measurements. Understanding the distinction is fundamental to everything else in this topic.

Forward modeling starts with a known (or assumed) subsurface model and predicts what geophysical data you'd observe. It requires a mathematical description of the physics linking subsurface properties to measurements. For example:

- Calculating gravity anomalies from a given density distribution
- Computing seismic travel times through a specified velocity model

Inverse modeling works in the opposite direction: you start with observed data and try to recover the subsurface model that produced it. The goal is to find a model whose predicted data (via forward modeling) matches the observations as closely as possible.

Challenges in Inverse Modeling

Inverse problems are typically ill-posed, which means they violate one or more of the conditions for a well-posed problem (existence, uniqueness, and stability of the solution). In practice, this shows up in two major ways:

- Non-uniqueness: Multiple subsurface models can explain the observed data equally well. You can't always tell which one is "correct."
- Instability: Small changes in the data (like noise) can produce large changes in the estimated model.

The relationship between model parameters and geophysical data is also often non-linear. A linear problem would let you solve for the model in one step using matrix algebra. Non-linearity means you typically need iterative optimization, where you repeatedly update your model estimate, run the forward problem, compare to data, and adjust.

Inversion Techniques for Subsurface Properties

Mathematical Formulation of Inversion

At its core, inversion is an optimization problem. You define an objective function that measures the misfit between predicted and observed data, then adjust model parameters to minimize it.

Common misfit measures include:

• Least-squares (L2-norm): $\phi = \sum(d_{\text{obs}} - d_{\text{pred}})^2$. Sensitive to outliers because large residuals get squared.

• L1-norm: $\phi = \sum|d_{\text{obs}} - d_{\text{pred}}|$. More robust to outliers since residuals aren't squared.

Because the problem is ill-posed, minimizing misfit alone isn't enough. Regularization adds a penalty term to the objective function that encodes your expectations about the model:

• Tikhonov regularization penalizes rough or complex models, favoring smooth solutions. The objective function becomes something like $\phi_{\text{total}} = \phi_{\text{data}} + \lambda\phi_{\text{reg}}$, where λ controls the trade-off between fitting the data and keeping the model smooth.

• Total variation regularization allows sharp boundaries in the model, which is useful when you expect distinct geological layers or contacts.

Optimization Algorithms and Probabilistic Inversion

Gradient-based methods are the workhorse of deterministic inversion. They iteratively update model parameters by computing how the objective function changes with respect to each parameter:

1. Start with an initial model guess.

2. Compute the gradient of the objective function (i.e., the direction of steepest increase in misfit).

3. Update the model in the direction that reduces misfit.

4. Repeat until convergence.

Two common variants:

• Steepest descent: Updates directly along the negative gradient. Simple but can converge slowly, especially near the solution.

• Conjugate gradient: Uses information from previous iterations to choose better search directions, improving convergence rate.

Markov chain Monte Carlo (MCMC) methods take a fundamentally different approach. Instead of seeking one best model, they sample the space of possible models to build up a posterior probability distribution:

1. Start with an initial model.

2. Propose a new model by perturbing the current one.

3. Evaluate whether the new model fits the data better (accounting for prior information).

4. Accept or reject the proposal based on a probability rule.

5. Repeat thousands to millions of times.

The resulting ensemble of accepted models maps out which parameter values are probable given the data. Common MCMC algorithms include the Metropolis-Hastings algorithm and the Gibbs sampler.

Limitations of Inversion Results

Non-Uniqueness and Resolution

No matter how good your inversion algorithm is, the results carry inherent limitations.

Non-uniqueness means multiple models can fit the data equally well. The data simply don't contain enough information to pin down a single answer. This is especially problematic when data coverage is sparse or when different physical properties produce similar signals.

Resolution describes how well the inversion can distinguish features at different scales and locations. It depends on:

- The spatial and temporal sampling density of your data
- The frequency content of the signal (higher frequencies resolve finer features)
- The physics of the measurement itself

The resolution matrix provides a formal way to assess this. In an ideal case, it would be an identity matrix (perfect resolution). In practice, off-diagonal elements show how features at one location "smear" into neighboring locations in the inverted model.

Uncertainty Quantification and Model Validation

Uncertainty in inverted models comes from several sources:

- Measurement errors (noise in the data)
- Modeling errors (simplified physics, missing geological complexity)
- Ill-posedness of the inverse problem itself

The model covariance matrix quantifies this uncertainty. Its diagonal elements give the variance of each model parameter (how uncertain that parameter is), while off-diagonal elements reveal correlations between parameters (when one parameter is poorly constrained, others linked to it may be as well).

Two key validation strategies:

- Sensitivity analysis: Perturb the input data or model parameters and observe how the inversion result changes. If small perturbations cause large changes, the result is poorly constrained.
- Cross-validation: Test whether the inverted model can predict data it wasn't trained on.
- Leave-one-out: Remove one data point, invert the rest, and check how well the result predicts the removed point.
- K-fold: Split data into K subsets, use each in turn as a validation set while inverting the remaining data.

Deterministic vs Probabilistic Inversion

Deterministic Inversion

Deterministic inversion seeks a single "best" model. The process follows these steps:

1. Choose an initial model and define an objective function.
2. Use a gradient-based algorithm to iteratively adjust model parameters.
3. Stop when a convergence criterion is met (e.g., misfit drops below a threshold) or a maximum number of iterations is reached.

The output is a point estimate of the subsurface properties. This is the most likely or optimal model given the data and your chosen objective function. The main trade-off: deterministic methods are computationally efficient, but they provide limited information about uncertainty. You get one answer, not a range of plausible answers.

Probabilistic Inversion

Probabilistic inversion estimates the full posterior probability distribution of model parameters, conditioned on the observed data and prior information. This follows from Bayes' theorem:

$$P(\mathbf{m}|\mathbf{d}) \propto P(\mathbf{d}|\mathbf{m}) \cdot P(\mathbf{m})$$

where $P(\mathbf{m}|\mathbf{d})$ is the posterior, $P(\mathbf{d}|\mathbf{m})$ is the likelihood, and $P(\mathbf{m})$ is the prior.

MCMC sampling is the most common approach. The ensemble of models generated by the chain represents the posterior distribution, from which you can extract:

- Marginal distributions for individual parameters
- Confidence intervals (e.g., the 95% credible interval for velocity at a given depth)
- Correlation structure between parameters

The trade-off is computational cost. Exploring a high-dimensional parameter space thoroughly requires many forward model evaluations.

Comparison and Hybrid Approaches

Deterministic: Computationally efficient, produces a single best-fit model. Best for large-scale problems or situations where computational resources are limited.

Probabilistic: Computationally expensive, but gives a full picture of uncertainty. Best when quantifying uncertainty matters for decision-making or risk assessment.

Hybrid approaches try to get the best of both worlds:

- Ensemble Kalman filtering maintains an ensemble of models (capturing uncertainty) but updates them sequentially as new data arrive, avoiding the full cost of MCMC.
- Particle swarm optimization uses a population of candidate solutions ("particles") that explore the parameter space collectively, balancing exploration of new regions with convergence toward promising solutions.

In practice, the choice between deterministic, probabilistic, and hybrid methods depends on the size of the problem, the computational budget, and how critical uncertainty estimates are for the application.

9.4 Integration of geophysical data sets

Integrating Geophysical Data Sets

Geophysical data integration combines multiple survey methods to build a more complete picture of subsurface structure and properties. By merging seismic, gravity, magnetic, and electrical data, you can resolve ambiguities that no single method could handle on its own. This is especially valuable in complex geological settings like fault zones, intrusive bodies, or mineral deposits, where one technique alone may miss critical features.

Benefits of Integrating Multiple Geophysical Data Sets

Each geophysical method is sensitive to different physical properties. Seismic data responds to acoustic impedance contrasts, gravity to density, magnetics to susceptibility, and electrical methods to conductivity. Combining them lets you leverage the strengths of each while compensating for their individual blind spots.

- Reduces non-uniqueness in interpretation by constraining models with complementary physical property information
- Improves spatial resolution, depth penetration, and feature sensitivity beyond what any single data set achieves
- Provides a more comprehensive subsurface characterization, particularly in heterogeneous settings where structure, lithology, and fluid content all vary simultaneously

Geophysical Data Sets Commonly Used in Integration

- Seismic data reveals subsurface structure and stratigraphy from the propagation and reflection of seismic waves (P-waves and S-waves). It excels at imaging layer boundaries and structural geometry.
- Gravity data maps variations in the Earth's gravitational field caused by lateral density contrasts. These anomalies highlight features like sedimentary basins (low density) or igneous intrusions (high density).
- Magnetic data detects variations in the Earth's magnetic field driven by the magnetic susceptibility of subsurface rocks, particularly useful for locating igneous bodies and concentrations of magnetic minerals like magnetite.
- Electrical and electromagnetic data measure the conductivity and resistivity of subsurface materials, making them sensitive to fluid content, salinity, and lithology. They're well suited for mapping aquifers, clay-rich zones, and ore deposits.

Each data set has its own resolution and depth capabilities. Integration works because the weaknesses of one method often align with the strengths of another.

Data Integration for Subsurface Characterization

Quantitative Data Integration Techniques

Joint inversion simultaneously inverts multiple geophysical data sets to produce a single subsurface model consistent with all the data. The process works as follows:

1. Define a common model parameterization that represents the subsurface in terms of physical properties relevant to all data sets (e.g., a shared grid of cells with density, velocity, and resistivity values).
2. Select a coupling strategy that links the different physical properties, either through empirical relationships (e.g., density-velocity correlations) or structural constraints (e.g., requiring property boundaries to coincide).
3. Construct a combined objective function that includes data misfit terms for each data set plus regularization terms.
4. Minimize the objective function iteratively, updating the model until it satisfies all data sets within their respective uncertainties.

Joint inversion exploits the complementary sensitivities of different methods, producing better-constrained models than inverting each data set independently.

Cooperative inversion takes a sequential approach: the result from inverting one data set is used to guide or constrain the inversion of another. For example, you might invert seismic data first to define structural boundaries, then use those boundaries as constraints when inverting gravity data. This ensures consistency between models derived from different methods without requiring a single unified objective function.

Qualitative Data Integration and Statistical Methods

Qualitative integration involves visually comparing and interpreting multiple data sets, often displayed as overlays in a GIS environment. This approach lets you identify spatial correlations and anomalies across data sets and integrate geophysical results with geological maps, geochemical surveys, and geotechnical logs.

Statistical methods add rigor to pattern recognition across data sets:

- Principal component analysis (PCA) reduces data dimensionality by identifying the components that explain the most variance. If you have five overlapping geophysical grids, PCA can highlight the dominant spatial patterns that cut across all of them, separating signal from noise.
- Cluster analysis groups spatially co-located data points based on their multi-attribute signatures. This helps delineate subsurface zones with distinct combinations of physical properties, even when no single attribute clearly defines the boundary.

Interpreting Integrated Geophysical Data

Multidisciplinary Approach to Interpretation

Interpreting integrated data requires synthesizing knowledge from geophysics, geology, and other relevant disciplines. The goal is to build a subsurface model that is internally consistent across all data sets and geologically plausible.

- Identify and delineate features such as lithological boundaries, faults, fractures, and zones of fluid or mineral accumulation (aquifers, hydrocarbon reservoirs, ore bodies)
- Account for the spatial and temporal resolution of each data set, recognizing that seismic data may resolve features at meter scale while gravity data operates at hundreds of meters
- Cross-check interpretations against known geological constraints to ensure the integrated model makes physical sense

Visualization and Uncertainty Analysis

Integrated models can be represented through several visualization approaches:

- Cross-sections provide 2D slices along vertical planes, showing spatial relationships between geological units and structures. These are useful for communicating structural interpretations along specific profiles.
- 3D models offer volumetric representations that allow interactive exploration of complex geological architectures. They're particularly valuable for understanding how structures connect in three dimensions.

• Attribute maps display the spatial distribution of specific properties (porosity, permeability, mineral content) derived from the integrated data, projected onto surfaces or depth slices.

Uncertainty analysis is essential for assessing model reliability:

1. Quantify the uncertainty in model parameters by propagating data errors and examining how well-constrained each part of the model is.

2. Apply techniques such as Monte Carlo simulation (running many model realizations with varied input parameters), sensitivity analysis (testing how changes in one parameter affect the result), and cross-validation (withholding subsets of data to test predictive accuracy).

3. Compare model predictions against independent observations, such as borehole data not used in the inversion, to validate accuracy.

Models that survive these tests carry more credibility for decision-making.

Benefits and Challenges of Integrated Data

Integration with Geological and Geotechnical Information

Geophysical data gains significant value when combined with direct subsurface measurements. These ground-truth data provide calibration points that anchor geophysical interpretations to real rock properties.

• Well logs supply continuous measurements of lithology, porosity, fluid content, and other properties along the borehole, bridging the scale gap between core samples and geophysical surveys.

• Core data provides physical rock and sediment samples for laboratory analysis, giving direct measurements of density, magnetic susceptibility, resistivity, and mechanical properties.

• Surface mapping documents the surface expression of geological structures and lithological units, providing boundary conditions for subsurface models.

This integration bridges the scale gap between large-scale geophysical measurements (which cover broad areas but with limited resolution) and small-scale geological observations (which are highly detailed but spatially sparse).

Challenges and Requirements for Effective Integration

Several practical challenges arise when combining diverse data sets:

• Scale mismatches: Geophysical data sets may have spatial resolutions ranging from meters (seismic) to kilometers (regional gravity), while well logs sample centimeter-scale intervals. Reconciling these scales requires careful upscaling or downscaling.

• Variable data quality: Uncertainty differs between data sets, so appropriate weighting and error propagation are needed during integration. Poorly constrained data should not dominate the result.

• Normalization and standardization: Different data types use different units, coordinate systems, and formats. These must be harmonized before meaningful integration can occur.

Effective integration also demands:

• Specialized workflows and software tools capable of handling diverse data formats in a unified framework

- Data management systems designed to store, organize, and retrieve multiple data sets along with their metadata
- Strong communication between geophysicists, geologists, and engineers throughout the project, from survey design through final interpretation

Applications and Decision-Making

Integrated subsurface models provide a more robust foundation for decision-making across a range of applications:

- Resource exploration and development: hydrocarbons, minerals, geothermal energy
- Geohazard assessment: earthquakes, landslides, subsidence
- Environmental management: groundwater contamination mapping, carbon sequestration site characterization
- Civil engineering: tunnel routing, dam siting, foundation design

By combining geophysical data with geological and geotechnical information, these models reduce the uncertainties inherent in subsurface work. The result is more informed planning, optimized survey design, and ultimately more cost-effective project execution.

Unit 10 – Geophysics in Resource Exploration

10.1 Petroleum geophysics and seismic exploration

Petroleum geophysics and seismic exploration are the primary tools used to find and develop oil and gas reservoirs. These methods use controlled sound waves to create detailed images of underground rock layers, allowing geophysicists to identify structures that may trap hydrocarbons.

Seismic data interpretation, combined with well log measurements, forms the backbone of modern exploration workflows. Understanding how these datasets work together is essential for mapping reservoirs, estimating reserves, and planning drilling operations.

Seismic Methods for Petroleum Exploration

Principles of Seismic Reflection and Refraction

Seismic reflection and refraction are the two foundational methods for imaging subsurface geology. Both rely on generating seismic waves with controlled sources (vibroseis trucks on land, air guns offshore, or explosives) and recording how those waves interact with rock layers below.

Seismic reflection occurs when waves hit an interface between two layers with different acoustic impedances and bounce back to the surface. Acoustic impedance is the product of a layer's density and seismic velocity ($Z = \rho \cdot v$). The greater the contrast in acoustic impedance between two layers, the stronger the reflection. This is what produces the visible horizons on a seismic section.

Seismic refraction occurs when waves encounter a layer with higher velocity, causing them to bend and travel along the interface before returning to the surface. The bending follows Snell's Law, which relates the angles of incidence and refraction to the velocities of the two layers:

$$\frac{\sin\theta_1}{v_1} = \frac{\sin\theta_2}{v_2}$$

Refraction is particularly useful for determining the velocity structure of the subsurface, which feeds into depth calculations.

Two-way travel time (TWT) is the time it takes for a seismic wave to travel from the source down to a reflector and back to the surface. Combined with a velocity model (built from refraction data and well logs), TWT is converted to actual depth, giving you the geometry of geological structures.

Applications and Advantages of Seismic Methods

Reflection seismic dominates petroleum exploration because it offers higher resolution and can image complex structures like faults, folds, and stratigraphic traps. High-resolution surveys can detect thin beds and subtle features down to several kilometers depth.

Refraction seismic plays a complementary role. It's most valuable for characterizing complex near-surface geology and providing velocity information used for static corrections and depth conversion of reflection data. It also sees use in deep crustal studies.

Both methods are non-invasive and can cover large areas efficiently, making them cost-effective compared to drilling exploratory wells. Advances in acquisition hardware, processing algorithms, and interpretation software continue to push the resolution and accuracy of subsurface imaging.

Interpreting Seismic Data for Reservoirs

Seismic Data Interpretation Techniques

Seismic sections are visual cross-sections of the subsurface, displaying two-way travel time on the vertical axis and distance along a survey line on the horizontal axis. They can be shown as wiggle traces, variable density plots, or color-coded displays.

The primary goal of interpretation is identifying hydrocarbon traps, which are geological configurations that accumulate and seal oil or gas. The three main trap types are:

- Anticlines: Upward-folded structures where hydrocarbons migrate to the crest and become trapped beneath a seal rock.
- Fault traps: Faults juxtapose permeable reservoir rock against impermeable rock, creating a lateral seal.
- Stratigraphic traps: Lateral changes in rock type or pinchouts of permeable layers create traps without structural deformation.

Certain patterns on seismic sections can directly indicate hydrocarbons:

- Bright spots: High-amplitude anomalies that may indicate gas or light oil in the pore space (the gas lowers acoustic impedance, increasing the reflection contrast).
- Flat spots: Horizontal reflections that cut across dipping structure, representing fluid contacts (gas-oil or oil-water boundaries).
- Dim spots / gas chimneys: Zones of reduced amplitude above a reservoir, caused by gas leaking upward and absorbing seismic energy.

Seismic attributes extract additional information from the data beyond simple reflection patterns:

- Amplitude attributes (e.g., RMS amplitude) highlight lithology or fluid changes.
- Frequency attributes (e.g., instantaneous frequency) can indicate variations in bed thickness or fluid type.
- Phase attributes help identify stratigraphic features and discontinuities.

Integrated Interpretation and Uncertainty Reduction

Seismic facies analysis classifies regions of a seismic section by their distinct reflection patterns, amplitude, frequency, and continuity. These facies can then be tied to depositional environments (channels, submarine fans, reefs) and lithologies (sandstone, shale, carbonate).

No single data type tells the full story. Interpretation improves significantly when seismic data is integrated with:

- Well logs: Detailed rock property and fluid measurements at specific borehole locations.
- Core analysis: Direct laboratory measurements of porosity, permeability, and fluid saturation from rock samples.
- Regional geological models: Broader structural and depositional context that guides and validates seismic picks.

Two advanced techniques are particularly important for reducing uncertainty:

1. Seismic inversion converts reflection data into quantitative rock properties (acoustic impedance or velocity), which can be directly compared with well log measurements.

2. AVO (Amplitude Versus Offset) analysis examines how reflection amplitude changes with source-receiver distance. Different fluids and lithologies produce characteristic AVO responses, making it a powerful tool for distinguishing gas-filled rock from brine-filled rock.

Reservoir Characterization with Logging and Attributes

Well Logging for Reservoir Properties

Well logging measures physical properties of rock formations along the length of a borehole, providing high-resolution vertical profiles of the reservoir. Logging tools are lowered into the well on a wireline cable, though modern techniques also include logging while drilling (LWD) and measurement while drilling (MWD), which record data in real time as the well is drilled.

The most common log types and what they measure:

Log Type	What It Measures	Primary Use
Gamma ray	Natural radioactivity	Distinguishing shale from clean reservoir rock
Density	Bulk density of formation	Porosity and lithology identification
Neutron porosity	Hydrogen content	Porosity estimation
Resistivity	Electrical resistivity	Fluid content and water saturation
Sonic	Compressional wave velocity	Porosity and mechanical properties

These logs are combined to determine the four key reservoir parameters:

• Lithology: Identified by cross-plotting gamma ray, density, and neutron logs.

• Porosity: Estimated from density, neutron, and sonic logs (often using two or more for cross-checking).

• Permeability: Estimated from porosity using empirical relationships (e.g., the Kozeny-Carman equation or core-calibrated transforms).

• Fluid saturation: Inferred primarily from resistivity logs using Archie's equation: $S_w = \left(\frac{a \cdot R_w}{\phi^m \cdot R_t}\right)^{1/n}$, where S_w is water saturation, R_w is formation water resistivity, R_t is true formation resistivity, ϕ is porosity, and a, m, n are empirical constants.

Seismic Attributes and Reservoir Characterization

Seismic attributes are quantitative values extracted from seismic trace data using mathematical algorithms. They can be computed along interpreted horizons, on time slices, or across entire 3D volumes.

Key attribute categories:

• Amplitude attributes (RMS amplitude, instantaneous amplitude): Highlight acoustic impedance contrasts tied to lithology or fluid changes.

• Frequency attributes (instantaneous frequency, dominant frequency): Sensitive to bed thickness variations and fluid type.

- Phase attributes (instantaneous phase, cosine of phase): Useful for identifying stratigraphic features and lateral discontinuities.
- Coherence attributes: Measure trace-to-trace similarity, making faults, fractures, and channel edges stand out clearly.
- Curvature attributes: Quantify the degree of folding or bending of reflectors, highlighting structural and stratigraphic features.

Seismic inversion deserves special attention because it bridges the gap between seismic data and rock properties. Two main approaches exist:

- Deterministic inversion: Uses a single starting model and produces one "best-fit" output of acoustic impedance or velocity.
- Stochastic inversion: Uses a range of input models and generates multiple equally probable realizations, capturing uncertainty in the result.

The real power comes from integrating well logs and seismic attributes together. Well logs give you high-resolution vertical detail at discrete points; seismic attributes give you spatially continuous coverage across the entire field. Geostatistical methods like kriging and co-kriging combine both data types to build 3D reservoir models that honor the well data while using seismic trends to interpolate between wells.

3D and 4D Seismic in Field Development

3D Seismic Surveys and Interpretation

3D seismic surveys acquire data across a dense grid of source and receiver lines, producing a full three-dimensional volume of the subsurface rather than isolated 2D cross-sections. The closely spaced lines provide high fold coverage (many overlapping recordings of the same subsurface point), which improves signal-to-noise ratio and image quality.

3D data enables far more accurate imaging of complex geology. 3D migration algorithms (Kirchhoff migration, wave-equation migration) reposition reflections to their true subsurface locations, correcting for the distortions that occur when structures are steeply dipping or laterally variable.

Three key interpretation techniques take advantage of the 3D volume:

1. Volume rendering: Displays the entire seismic cube as a semi-transparent volume, letting you visualize spatial relationships between faults, channels, and other features from any angle.
2. Horizon slicing: Extracts seismic attributes along a picked horizon surface, mapping lateral variations in reservoir properties across the field.
3. Attribute analysis on volumes: Applies coherence, curvature, or amplitude attributes across the full 3D dataset to highlight faults, channel systems, or fluid contacts that might not be visible on individual sections.

4D Seismic Monitoring and Reservoir Management

4D seismic (time-lapse seismic) repeats a 3D survey over the same area at different times during production, typically at intervals of several years. The surveys must be processed carefully to ensure consistency, so that any differences between them reflect actual reservoir changes rather than acquisition artifacts.

4D seismic detects production-related changes in the reservoir:

- Fluid substitution: As water replaces oil or gas during production, acoustic impedance changes, producing a detectable shift in seismic response.
- Pressure depletion: Reduced pore pressure can cause compaction and velocity changes, altering travel times.
- Bypassed reserves: Areas that haven't been swept by the production process show up as unchanged zones, guiding decisions about where to drill infill wells.

Two primary interpretation techniques are used with 4D data:

1. Difference volumes: Subtract the baseline (pre-production) survey from the monitor (later) survey. Any non-zero signal highlights where the reservoir has changed.
2. Time-shift analysis: Measures travel-time differences between baseline and monitor surveys. These shifts indicate velocity changes caused by pressure depletion or fluid substitution.

Integrating 3D and 4D seismic with reservoir simulation models closes the loop between observation and prediction. Simulation models use seismic-derived properties (porosity, permeability) to predict fluid flow and production. When 4D seismic reveals discrepancies between predicted and actual reservoir behavior, the model can be calibrated and updated. This integrated workflow, combining seismic, well, and production data, enables data-driven decisions about well placement, injection strategy, and overall field management.

10.2 Mineral exploration geophysics

Mineral exploration geophysics applies physical measurements at the Earth's surface and in boreholes to detect subsurface ore bodies and alteration systems. Because most economic mineral deposits differ from their host rocks in density, magnetism, conductivity, or radioactivity, geophysical surveys can narrow vast search areas down to drill-ready targets. This guide covers the main survey methods, how to interpret the anomalies they produce, how different datasets are integrated, and the role of borehole geophysics.

Geophysical Methods for Mineral Exploration

Gravity, Magnetic, and Electromagnetic Surveys

Gravity surveys measure small variations in the Earth's gravitational field caused by lateral density contrasts in the subsurface. A massive sulfide lens (density ~4.5 g/cm^3) sitting in felsic volcanic rock (~2.7 g/cm^3) produces a measurable positive gravity anomaly. After applying standard corrections (free-air, Bouguer, terrain), the residual anomaly map highlights density contrasts that may correspond to ore bodies.

Magnetic surveys map variations in the total magnetic field intensity. The primary target mineral is magnetite, but pyrrhotite and other ferrimagnetic phases also contribute. Magnetic data are especially useful for tracing iron-oxide-rich alteration halos around deposits and for mapping subsurface geology under cover. One complication is remanent magnetization: magnetic minerals can retain a magnetization direction acquired during formation that differs from the present-day field, causing anomalies that don't align with the expected response.

Electromagnetic (EM) surveys detect electrically conductive or magnetically susceptible bodies by measuring how they interact with a transmitted EM field.

- Frequency-domain EM methods, such as controlled-source audio-frequency magnetotellurics (CSAMT), transmit a range of frequencies. Lower frequencies penetrate deeper, so sweeping across frequencies builds a conductivity profile with depth.

- Time-domain EM (TDEM or TEM) methods transmit a pulse and then measure the decaying secondary field. The decay rate and shape reveal the conductivity, size, and depth of a target. Massive sulfide bodies, for example, produce a slow, strong decay compared to weakly conductive overburden.

Induced Polarization and Radiometric Surveys

Induced polarization (IP) measures the chargeability of subsurface materials. When current is injected into the ground and then switched off, certain minerals cause the voltage to decay slowly rather than dropping to zero instantly. Disseminated sulfide grains (chalcopyrite, pyrite) are the classic source of this effect. IP is particularly valuable for porphyry copper and epithermal gold systems, where sulfides are finely disseminated through large rock volumes and may not produce strong gravity, magnetic, or EM anomalies on their own.

Radiometric surveys measure natural gamma radiation from the decay of potassium (^{40}K), uranium (^{238}U), and thorium (^{232}Th). These surveys are most effective in areas with thin or no overburden. They can identify:

- Potassic alteration zones (elevated K), which commonly surround porphyry copper deposits
- Uranium mineralization in sedimentary basins, where U anomalies directly indicate the target commodity
- Lithological boundaries, since different rock types have distinct K-U-Th signatures

Together, IP and radiometric surveys fill gaps left by gravity, magnetic, and EM methods, adding information about disseminated mineralization and alteration chemistry.

Interpretation of Geophysical Data

Anomalies and Their Significance

Every geophysical method produces anomalies, but not every anomaly means ore. The table below summarizes common anomaly sources and the ambiguities involved:

Method	Anomaly type	Possible ore-related source	Possible non-ore source
Gravity	Positive residual	Massive sulfide, IOCG deposit	Mafic intrusion, dense basement
Magnetic	High / low dipolar	Magnetite in skarn or BIF	Mafic dyke, remanent magnetization artifact
EM	Late-time conductor	Massive sulfide lens	Graphitic shale, saline groundwater
IP	High chargeability	Disseminated chalcopyrite/pyrite	Graphite, clay-rich zones, magnetite
Radiometric	Elevated K	Potassic alteration halo	K-feldspar-rich granite, weathering effects

The key takeaway: a single anomaly type is rarely diagnostic on its own. Reliable interpretation requires stacking multiple lines of evidence.

Geological Context and Interpretation

Geophysical data should always be interpreted within the geological framework. A strong EM conductor in a greenstone belt terrane is a high-priority massive sulfide target; the same conductor in a sedimentary basin may just be a graphitic shale horizon.

Steps for sound interpretation:

1. Establish the geological setting from mapping, stratigraphy, and known structural controls.
2. Identify coherent anomalies that stand out from regional background trends.
3. Cross-check anomalies across methods. A coincident gravity high, EM conductor, and IP chargeability high is far more compelling than any one alone.
4. Apply geophysical modeling. 2D and 3D inversion algorithms estimate the depth, geometry, and physical properties (density, susceptibility, conductivity) of the body causing the anomaly. These models are non-unique, so geological constraints are essential to guide them toward realistic solutions.
5. Rank targets for follow-up (infill surveys, drilling) based on the strength and coherence of multi-method anomalies and their geological plausibility.

Integration of Geophysical Data

Combining Geophysical, Geological, and Geochemical Information

No single dataset tells the whole story. Effective mineral exploration integrates several types of information:

- Geological mapping provides lithological, stratigraphic, and structural context. It tells you where certain deposit types are geologically permissible, so you don't chase anomalies in the wrong rock package.
- Geochemical sampling (soil surveys, stream sediment surveys, rock chip analyses) highlights areas with elevated pathfinder elements (e.g., arsenic and antimony around gold deposits). A geophysical anomaly that coincides with a geochemical anomaly jumps up the priority list.
- Petrophysical measurements (density, magnetic susceptibility, resistivity, chargeability) on rock samples and drill core provide the physical property link between geology and geophysical response. Without petrophysical data, modeling is poorly constrained.
- Geochronological data (U-Pb, Ar-Ar dating of intrusions and alteration minerals) establish the timing of mineralization events, helping you determine whether a geophysical anomaly is associated with the right-aged system.

Data Integration Techniques and Tools

- GIS platforms allow you to overlay geophysical grids, geological maps, geochemical point data, and structural interpretations in a common spatial framework. Simple overlay analysis can quickly highlight areas where multiple favorable indicators converge.
- 3D modeling software (e.g., Leapfrog, GOCAD) builds volumetric models that incorporate geophysical inversions, geological surfaces, and drill hole data. These models let you visualize how an anomaly relates to known geology at depth.

- Machine learning approaches, including unsupervised clustering and supervised classification, can identify subtle multi-variable patterns across large datasets that a human analyst might miss. These are tools to assist interpretation, not replace geological reasoning.

The goal of integration is to move from a large number of possible targets to a short list of well-supported drill targets, reducing exploration risk and cost.

Borehole Geophysics in Mineral Exploration

In-Situ Measurements and Characterization

Once drilling begins, borehole geophysical logging provides continuous, in-situ measurements of physical properties along the hole. These logs serve two purposes: they characterize the intersected geology directly, and they supply ground-truth data for calibrating surface geophysical interpretations.

Common borehole logging tools in mineral exploration:

- Gamma-gamma density: Measures bulk density by detecting back-scattered gamma rays. Essential for resource estimation (converting volume to tonnage) and for calibrating surface gravity models.

- Magnetic susceptibility: Identifies magnetic mineral distribution downhole. Particularly useful where remanent magnetization complicates surface magnetic interpretation.

- Resistivity and IP logs: Delineate conductive or chargeable zones tied to sulfide mineralization or alteration, at much higher spatial resolution than surface surveys.

- Natural gamma: Detects K, U, and Th concentrations, identifying potassic alteration or uranium-bearing intervals.

Structural Analysis and Resource Estimation

Acoustic televiewer (ATV) and optical televiewer (OTV) tools produce oriented images of the borehole wall. From these images you can pick fractures, veins, lithological contacts, and faults with their true orientation (dip and dip direction). This structural information is critical when mineralization is structurally controlled, because it tells you which direction to drill next.

Borehole geophysical data feed back into the broader exploration program in several ways:

1. Calibrate surface models. Density and susceptibility logs constrain inversion models, reducing ambiguity.

2. Refine resource estimates. Continuous physical property logs fill gaps between assayed intervals and improve geological domaining.

3. Guide future drilling. Structural data from televiewers, combined with updated 3D models, help optimize the location and orientation of the next holes, targeting extensions of known mineralization or testing new anomalies with greater confidence.

10.3 Groundwater exploration and hydrogeophysics

Electrical Resistivity and Electromagnetic Methods for Groundwater Exploration

Groundwater exploration depends on geophysical methods to map subsurface structures and locate water resources without drilling. Electrical resistivity, electromagnetic surveys, seismic techniques, and ground-penetrating radar each reveal different aquifer properties and groundwater dynamics. These same methods are also critical for tracking contaminant plumes and evaluating remediation efforts.

Principles of Electrical Resistivity Methods

Electrical resistivity methods work by injecting current into the ground through electrodes and measuring the resulting voltage differences. The measured resistance depends on lithology, porosity, and water content of subsurface materials.

The resistivity of rocks and sediments is primarily controlled by the amount and salinity of pore water. This relationship is described by Archie's Law, which relates bulk resistivity to porosity, water saturation, and pore fluid resistivity. In practice, this means:

- Saturated zones exhibit lower resistivity than unsaturated zones because pore water conducts current far better than rock matrix or air.
- Saline groundwater drives resistivity even lower, while fresh groundwater in clean sands may still show moderate resistivity.
- Clay-rich materials also reduce resistivity due to surface conduction along clay mineral surfaces, independent of pore water salinity.

Resistivity surveys use various electrode configurations, each with trade-offs:

- Wenner array: Good vertical resolution, straightforward field setup, but relatively low lateral resolution.
- Schlumberger array: Better depth sounding capability with fewer electrode moves; commonly used for vertical electrical sounding (VES).
- Dipole-dipole array: Strong lateral resolution, making it well-suited for 2D imaging, but more sensitive to noise at large separations.

Increasing the electrode spacing increases the depth of investigation, but at the cost of reduced resolution.

Principles of Electromagnetic Methods

Electromagnetic (EM) methods induce electromagnetic fields in the subsurface and measure the secondary response. Unlike resistivity methods, EM techniques don't require galvanic contact with the ground, which makes them faster to deploy over large areas.

Two main categories exist:

- Frequency-domain EM (FDEM): Transmits a continuous signal at one or more frequencies. Lower frequencies penetrate deeper. Well-suited for rapid mapping of lateral conductivity variations.
- Time-domain EM (TDEM): Transmits a pulse and measures the decay of the secondary field over time. Later time gates sample deeper. Particularly effective for detecting conductive targets at depth.

EM methods are especially sensitive to conductive materials. Clay-rich sediments and saline groundwater produce strong responses, making EM useful for mapping aquitards, saltwater intrusion zones, and the boundaries between fresh and saline water. The depth of investigation depends on transmitter-receiver separation (FDEM), measurement time window (TDEM), and the conductivity structure itself: highly conductive near-surface layers limit penetration into deeper targets.

Interpreting Geophysical Data for Groundwater Resources

Inverting Geophysical Data for Subsurface Models

Raw geophysical measurements (apparent resistivity, EM voltages) don't directly show you the subsurface structure. They need to be inverted to produce models of true resistivity or conductivity distribution. The inversion process works in these steps:

1. Collect field data along profiles or in grid patterns.
2. Define an initial subsurface model (a grid of cells, each with an assigned resistivity).
3. Use forward modeling to predict what the data would look like for that model.
4. Compare predicted data to observed data and calculate the misfit.
5. Iteratively adjust the model to minimize the misfit, subject to regularization constraints that keep the model geologically reasonable.
6. Evaluate the final model for resolution and reliability, particularly at depth where sensitivity decreases.

The resulting 2D or 3D resistivity models can then be interpreted in hydrogeological terms:

- Low-resistivity zones (typically < 10–50 Ω·m, depending on geology) may indicate saturated, permeable aquifers or clay-rich units.
- High-resistivity zones (hundreds to thousands of Ω·m) often represent unsaturated material, clean bedrock, or low-permeability formations.
- Conductive anomalies in EM data could mean clay-rich aquitards or saline groundwater. Distinguishing between these requires additional information.

This ambiguity is a key challenge: a low-resistivity anomaly could be a productive freshwater aquifer in saturated sand, a clay aquitard, or a saline zone. You can't tell from resistivity alone.

Integrating Geophysical Data with Other Information

Because of the non-uniqueness problem described above, geophysical data should always be integrated with:

- Borehole logs (lithological descriptions, geophysical well logs) to calibrate resistivity values against known geology.
- Hydraulic test data (pumping tests, slug tests) to relate geophysical models to actual hydraulic conductivity and transmissivity.
- Water quality data to distinguish salinity-driven conductivity from clay-driven conductivity.

Time-lapse geophysical surveys add a temporal dimension. By repeating measurements over weeks, months, or seasons, you can monitor:

- Changes in groundwater levels (the water table boundary shifts in resistivity images).
- Seasonal or pumping-induced salinity changes.
- Aquifer storage variations, which inform sustainability assessments.

Time-lapse approaches are powerful because even if the absolute resistivity is ambiguous, changes in resistivity over time are directly linked to changes in saturation or water chemistry.

Seismic and Ground-Penetrating Radar in Hydrogeological Studies

Seismic Methods for Hydrogeological Characterization

Seismic methods measure how elastic waves propagate through the subsurface. Wave velocities depend on the density and elastic moduli of rocks and sediments, which in turn relate to lithology, porosity, and saturation.

Two main approaches are used in hydrogeological work:

- Seismic refraction: Measures the travel times of waves refracted along layer boundaries. Particularly useful for mapping the depth to bedrock or the water table, because seismic velocity increases sharply at these interfaces. P-wave velocity jumps from roughly 200–800 m/s in unsaturated sediments to 1,500+ m/s below the water table (since water is nearly incompressible).
- Seismic reflection: Records waves reflected from subsurface interfaces. Provides detailed images of stratigraphy, bedrock topography, and structural features like faults or buried channels that may control groundwater flow paths.

Seismic methods complement resistivity and EM surveys because they respond to mechanical properties rather than electrical properties. A clay layer and a sand aquifer might have similar seismic velocities but very different resistivities, so combining both datasets reduces interpretation ambiguity.

Ground-Penetrating Radar for Shallow Subsurface Imaging

Ground-penetrating radar (GPR) transmits high-frequency electromagnetic pulses (typically 10 MHz to 1 GHz) into the ground and records reflections from boundaries where the dielectric permittivity changes. The dielectric permittivity of a material is strongly influenced by water content, since water has a much higher permittivity ($\varepsilon_r \approx 80$) than most minerals ($\varepsilon_r \approx 4\text{–}8$) or air ($\varepsilon_r = 1$).

GPR is effective for mapping:

- Shallow aquifer geometry and the water table surface.
- Bedrock surfaces and sedimentary structures (e.g., channel fills, gravel lenses).
- Subsurface utilities, cavities, or buried infrastructure that may affect groundwater flow.

The trade-off with GPR is between resolution and penetration depth. Higher antenna frequencies give finer resolution but attenuate faster. In conductive materials (clay-rich soils, saline groundwater), GPR signal is rapidly absorbed, limiting penetration to less than a meter in some cases. In resistive materials (dry sand, gravel, limestone), penetration can reach 10–30 m or more.

GPR works best in resistive, low-clay environments. If the site has significant clay content, GPR may not be the right tool, and EM or resistivity methods will likely perform better.

Geophysics in Contaminant Transport and Remediation

Characterizing Subsurface Properties for Contaminant Transport

Contaminant transport through the subsurface is governed by porosity, permeability, and hydraulic conductivity. Geophysical methods can characterize these properties indirectly:

- Electrical resistivity and EM surveys detect conductive contaminant plumes. Leachate from landfills, industrial effluent, or road salt runoff typically has elevated dissolved solids, making the plume more conductive than the surrounding groundwater. Repeated surveys can map the extent and migration direction of the plume over time.

- Seismic and GPR methods identify preferential flow pathways such as fracture zones, faults, or high-permeability sand and gravel channels. These features can cause contaminants to travel much faster and farther than predicted by simple uniform-flow models.

This information directly guides practical decisions: where to place monitoring wells, how to orient extraction wells, and where barriers or reactive zones should be installed.

Monitoring Remediation Efforts with Geophysical Methods

Time-lapse geophysical monitoring is one of the most valuable applications in remediation. Rather than relying solely on point measurements from monitoring wells, geophysical surveys provide spatially continuous information about what's happening between wells.

Specific applications include:

- Pump-and-treat systems: Resistivity monitoring can track whether the conductive plume is shrinking as contaminated water is extracted and treated.

- In-situ bioremediation: Microbial activity changes pore water chemistry (pH, dissolved gases, ionic strength), which can alter the bulk resistivity. Time-lapse surveys can map where bioremediation is active.

- Chemical oxidation/reduction: Injected amendments (permanganate, zero-valent iron, etc.) change the electrical properties of the subsurface. Geophysical monitoring can track the distribution of injected reagents and verify they've reached the target zone.

Integrating geophysical monitoring data with numerical hydrogeological models improves predictions of contaminant fate and transport. This feedback loop supports adaptive management: if the geophysical data show the plume migrating in an unexpected direction, the remediation strategy can be adjusted accordingly.

10.4 Environmental geophysics and site characterization

Geophysical Methods for Site Characterization

Environmental geophysics applies non-invasive survey techniques to map subsurface conditions at contaminated or development sites. These methods let you detect buried objects, track contaminant plumes, and characterize soil and rock properties without extensive drilling. The results feed directly into site conceptual models that guide remediation design and long-term monitoring.

Electrical Resistivity Methods

Electrical resistivity methods measure how strongly subsurface materials resist the flow of electric current. Resistivity values vary with lithology (clay vs. sand vs. bedrock), porosity, and the type of fluid filling pore spaces (fresh water, saline water, or contaminants). This makes resistivity a powerful tool for distinguishing geological layers and spotting anomalies.

Two key techniques fall under this heading:

- DC resistivity injects current into the ground through surface electrodes and measures the resulting potential difference. By varying electrode spacing and geometry, you build up a profile of how resistivity changes with depth and lateral position.

- Induced polarization (IP) measures the voltage decay after current injection is switched off. Materials like clay minerals or metallic sulfides (pyrite, galena) store and release charge slowly, producing a measurable IP response. This helps distinguish clay-rich zones or mineralized contamination from clean sand or gravel.

Ground-Penetrating Radar (GPR)

Ground-penetrating radar transmits high-frequency electromagnetic pulses into the ground and records reflections from subsurface interfaces. Reflections occur wherever there's a contrast in dielectric properties, which are controlled mainly by water content, lithology, and the presence of man-made objects (pipes, tanks, foundations).

Two trade-offs govern GPR performance:

- Higher antenna frequencies (e.g., 400–1600 MHz) give better resolution but shallower penetration.

- Lower frequencies (e.g., 25–200 MHz) penetrate deeper but resolve less detail.

Electrically conductive materials (clay-rich soils, saline groundwater) strongly attenuate GPR signals, limiting depth of investigation in those settings.

Other Methods

Several additional techniques see regular use in environmental site characterization:

- Seismic refraction and reflection determine bedrock depth, sedimentary layering, and water table position.

- Electromagnetic (EM) induction maps lateral variations in ground conductivity without requiring ground contact, making it efficient for large-area surveys.

- Magnetic surveys detect ferrous metallic objects like buried drums, steel tanks, or pipelines.

The choice of method depends on target depth, required resolution, site conditions (terrain, vegetation, cultural noise), and the physical property contrasts you're trying to exploit.

Interpreting Geophysical Data

Electrical Resistivity and IP Data Interpretation

Raw resistivity measurements are inverted to produce 2D or 3D models of the subsurface resistivity distribution. These models can reveal:

- Lithological boundaries (e.g., a clay aquitard overlying sand)

- Fracture zones (which often appear as low-resistivity linear features)
- Groundwater salinity variations (saltwater is far more conductive than freshwater)

IP data add another diagnostic layer. Elevated IP responses can indicate clay-rich horizons or contamination plumes containing polarizable materials such as hydrocarbons or dissolved heavy metals (lead, chromium). Combining resistivity and IP results helps you distinguish, for example, a clean clay layer from a contaminated sand unit that might have similar resistivity values.

GPR and Seismic Data Interpretation

GPR profiles display reflected wave travel times, which you convert to depth using the electromagnetic wave velocity of the subsurface materials (typically estimated from hyperbola fitting or common-midpoint surveys).

- Continuous reflectors delineate stratigraphic layers, the water table, or the tops of buried structures.
- Discrete hyperbolic reflections often indicate point targets like pipes or boulders.
- Signal attenuation (loss of reflections at depth) can itself be diagnostic, pointing to conductive materials such as clay or contaminated groundwater.

Seismic data reveal bedrock geometry, sedimentary layering, and the water table through velocity contrasts. They can also identify voids, fracture networks, and low-velocity zones caused by saturated unconsolidated fill.

Electromagnetic and Magnetic Data Interpretation

EM induction maps apparent conductivity across a site, highlighting conductive anomalies like leachate plumes or zones of saline intrusion. Magnetic surveys detect ferromagnetic objects (buried drums, steel pipes, rebar) as dipolar anomalies. Together, these methods are especially useful for rapid screening of large sites before deploying more detailed techniques like resistivity or GPR.

Integrating Geophysical Data for Site Assessment

Data Integration Techniques

No single data type tells the whole story. Effective site characterization combines three streams of information:

- Geophysical data provide spatially continuous images of subsurface properties (resistivity, velocity, dielectric constant).
- Geotechnical data (borehole logs, soil index properties, penetration tests) supply ground-truth constraints at discrete locations.
- Geochemical data (contaminant concentrations, pH, redox potential) characterize the nature and extent of contamination and help explain geophysical anomalies.

Bringing these together produces a site conceptual model that identifies contaminant sources (leaking tanks, surface spills), transport pathways (fractures, permeable sand layers), and receptors (drinking water wells, surface water bodies).

Common integration approaches include:

1. Co-located data comparison — overlaying resistivity profiles on borehole logs to calibrate geophysical interpretations.

2. Geostatistical methods — using techniques like kriging to interpolate between sparse borehole data, guided by the denser geophysical coverage.

3. Joint or coupled inversion — simultaneously inverting multiple geophysical datasets (e.g., resistivity and seismic) to produce models that satisfy both datasets and reduce ambiguity.

Applications in Site Assessment and Remediation Planning

The integrated site model directly supports practical decisions: where to place monitoring wells, how to orient extraction or injection wells, and where to install barriers or caps. By reducing uncertainty about subsurface conditions, geophysical integration cuts the number of boreholes needed and helps avoid costly surprises during construction.

Geophysics in Subsurface Monitoring and Remediation

Monitoring Subsurface Processes

Time-lapse geophysical surveys repeat measurements at the same locations over weeks, months, or years to track changes in subsurface conditions. The baseline survey establishes a reference, and subsequent surveys reveal how properties evolve.

Specific monitoring applications include:

- Resistivity and IP monitoring can track the migration of conductive contaminant plumes or detect geochemical changes caused by remediation activities like biostimulation or in-situ chemical oxidation.
- Repeat GPR surveys can detect changes in the saturation and spatial distribution of non-aqueous phase liquids (NAPLs) during pump-and-treat or surfactant flushing, and can verify the integrity of containment structures (slurry walls, engineered caps).
- Seismic monitoring can detect changes in mechanical properties, such as fracture development during hydraulic treatment or soil consolidation following stabilization.

Assessing Remediation Effectiveness

Time-lapse geophysical data serve two roles in evaluating remediation:

1. Model calibration and validation — Monitoring results can be compared against numerical flow-and-transport models to test predictions and refine parameters, improving forecasts of how the site will evolve.

2. Performance assessment — Comparing current surveys to the baseline lets you quantify progress toward cleanup goals. For example, increasing resistivity in a previously conductive plume zone suggests declining contaminant concentrations, while reduced GPR attenuation may indicate NAPL removal.

These comparisons help site managers decide whether to continue, modify, or conclude active remediation.

Unit 11 – Geophysical Hazards and Risk Assessment

11.1 Seismic hazard assessment and earthquake prediction

Seismic hazard assessment and earthquake prediction are central to understanding and reducing earthquake risks. These methods draw on statistical analysis, historical data, and geological evidence to estimate the likelihood and potential impact of future earthquakes at specific sites.

Short-term earthquake prediction remains one of the hardest problems in geophysics, but long-term forecasting is well-established and directly shapes building codes, land-use planning, and emergency preparedness. Seismic hazard maps translate all of this analysis into practical tools that guide decision-making in earthquake-prone regions.

Probabilistic Seismic Hazard Assessment

Key Components and Methods

Probabilistic Seismic Hazard Assessment (PSHA) is a statistical framework that quantifies the probability of exceeding a certain ground motion level at a specific site over a given time period. It's the backbone of modern seismic hazard analysis, and it combines several components into a single, integrated result.

PSHA follows a general workflow:

1. Seismic source characterization — Identify and model potential earthquake sources, including their geometry (fault length, dip, depth), maximum magnitude (M_{max}), and recurrence rates.
2. Ground motion prediction equations (GMPEs) — Apply attenuation relationships that describe how ground motion decreases with distance from the source. GMPEs account for magnitude, source-to-site distance, and local site conditions (soil type, bedrock depth, basin effects).
3. Hazard curve calculation — For each site, compute the annual probability of exceeding various levels of ground motion (typically peak ground acceleration, PGA, or spectral acceleration, S_a).
4. Hazard map generation — Display the spatial distribution of ground motion levels for a chosen probability of exceedance. Common reference levels are 2% probability of exceedance in 50 years (roughly a 2,475-year return period) and 10% in 50 years (475-year return period).

The distinction between hazard curves and hazard maps matters: a hazard curve shows the full range of exceedance probabilities for a single site, while a hazard map shows one exceedance probability across an entire region.

Accounting for Uncertainties

A major strength of PSHA is its formal treatment of uncertainty. Two types matter here:

- Epistemic uncertainty arises from incomplete knowledge. Examples include poorly constrained fault geometries, uncertain M_{max} estimates, and limited strong-motion datasets used to calibrate GMPEs.
- Aleatory variability reflects the natural randomness in earthquake processes, such as the scatter in recorded ground motions even for events of similar magnitude and distance.

Logic trees are the standard tool for handling epistemic uncertainty. Each branch of the tree represents an alternative model or parameter choice (e.g., different GMPEs, different fault slip rates), and each branch is assigned a weight reflecting its relative credibility. The final hazard result integrates across all branches, capturing the full range of possible outcomes rather than relying on a single "best guess."

By combining logic trees with the inherent aleatory variability in GMPEs, PSHA produces hazard estimates that are more robust and transparent about what we know and what we don't.

Historical Seismicity Data for Hazard Assessment

Sources and Applications of Historical Data

Reliable earthquake records are the foundation of any hazard assessment. These records come from two main sources:

- Instrumental data, recorded by seismometers since the late 19th century (and with good global coverage since the mid-20th century), provide precise locations, magnitudes, depths, and focal mechanisms.
- Historical accounts, including written records, newspaper reports, and personal diaries, extend the record further back in time. These are especially valuable in regions with long written histories (e.g., China, the Mediterranean, Japan) but require careful calibration to assign approximate magnitudes and locations.

Both types of data feed into earthquake catalogs, which compile events for a region and form the basis for estimating seismicity rates. A key relationship derived from these catalogs is the Gutenberg-Richter law:

$$\log_{10} N = a - bM$$

where N is the number of earthquakes with magnitude $\geq M$, a describes the overall rate of seismicity, and b (typically close to 1.0) describes the relative proportion of large versus small events. This relationship is fundamental to estimating recurrence rates in PSHA.

Paleoseismic Evidence

Instrumental and historical records cover only a few hundred years at best, which is far too short to capture the full behavior of faults with recurrence intervals of thousands of years. Paleoseismology extends the earthquake record into the geologic past using physical evidence preserved in the landscape and subsurface.

Key methods include:

- Fault trenching — Excavating across an active fault to expose displaced and deformed sedimentary layers. Each disrupted horizon can represent a past earthquake, and dating techniques (radiocarbon, optically stimulated luminescence) constrain when each event occurred.
- Liquefaction features — Sand blows and sand dikes form when strong shaking causes saturated, loose sediments to lose strength and erupt to the surface. Finding and dating these features in the stratigraphic record provides evidence of past strong shaking even where no surface fault rupture occurred.
- Offset geomorphic features — Displaced stream channels, offset terrace risers, and warped shorelines record cumulative fault slip. Dating these offsets with radiocarbon or cosmogenic nuclide methods (e.g., ^{10}Be, ^{26}Al) yields slip rates and recurrence intervals.

Paleoseismic data are critical for constraining M_{max} and recurrence intervals on specific faults, directly feeding into the seismic source characterization step of PSHA.

Limitations of Earthquake Prediction

Challenges in Short-term Prediction

There's an important distinction to keep straight: prediction means specifying the location, time window, and magnitude of a future earthquake with enough precision to be actionable. Forecasting means providing probabilities of future seismic events over a broader time window and area.

Short-term prediction remains largely unachieved, for several reasons:

- The physics of earthquake nucleation (how a rupture initiates on a fault) is still not well understood. We don't have a reliable physical model that tells us when accumulated stress will be released.
- Proposed precursory phenomena have not proven consistent or reliable. These include foreshock sequences, changes in seismicity patterns (seismic quiescence or activation), groundwater level fluctuations, radon gas emissions, and electromagnetic anomalies. Some of these have been observed before certain earthquakes but are absent before others, and they also occur without any subsequent earthquake.
- The chaotic, nonlinear nature of fault systems means that small, unmeasurable differences in stress state can determine whether a fault ruptures now or decades from now.

The 1975 Haicheng earthquake in China is often cited as a successful prediction (based on foreshock activity), but the 1976 Tangshan earthquake, which killed over 240,000 people, came with no recognized precursors. This contrast illustrates the fundamental unreliability of current precursor-based approaches.

Long-term Forecasting and Consequences

Long-term forecasting is more tractable and forms the practical basis for hazard mitigation. Two common approaches:

- Time-independent models use average recurrence rates (from Gutenberg-Richter statistics) and assume earthquakes are equally likely at any time. These are the simplest and most widely used in PSHA.
- Time-dependent models (renewal models, stress transfer models) account for the time elapsed since the last major event on a fault. The idea is that probability increases as stress re-accumulates. Coulomb stress transfer models also consider how a nearby earthquake may have brought a fault closer to (or further from) failure.

Even with these tools, forecasts carry substantial uncertainty. The societal stakes of getting it wrong are high:

- False alarms can trigger unnecessary evacuations, economic disruption, and erosion of public trust in scientific institutions.
- Missed events can lead to catastrophic losses if communities are unprepared.

Communicating probabilistic forecasts to the public and to policymakers is itself a major challenge. Probabilities like "a 62% chance of a $M \geq 6.7$ earthquake in the San Francisco Bay Area within 30 years" (from the USGS 2007 Working Group) are meaningful to scientists but can be difficult for non-specialists to interpret and act on.

Seismic Hazard Maps for Risk Mitigation

Applications in Building Codes and Land-use Planning

Seismic hazard maps are the primary way PSHA results reach engineers, planners, and policymakers. They display expected ground motion levels (usually PGA or spectral acceleration at specific periods) for a chosen probability of exceedance across a region.

Their applications include:

- Building codes — Modern risk-targeted codes (e.g., ASCE 7 in the United States, Eurocode 8 in Europe) use hazard map values to set minimum design requirements for structures. A building in Los Angeles faces different design ground motions than one in Houston, and the hazard map quantifies that difference.
- Seismic retrofitting — Hazard maps help prioritize which existing buildings and infrastructure need strengthening first, focusing resources on the highest-hazard areas.
- Land-use planning — Zoning regulations can steer critical facilities (hospitals, schools, fire stations) and lifelines (bridges, water mains, power grids) away from the most hazardous zones, including areas prone to surface rupture, liquefaction, or landslides.

Risk Assessment and Emergency Response Planning

Beyond engineering design, hazard maps support broader risk management:

- Insurance — Companies use seismic hazard data to assess portfolio risk and set premiums for earthquake insurance. Higher-hazard zones correspond to higher premiums.
- Emergency response planning — Hazard maps inform the design of evacuation routes, placement of shelters and medical facilities, and pre-positioning of supplies for post-earthquake response.
- Scenario modeling — Scenario-based maps simulate ground motions and impacts from specific plausible earthquakes (e.g., a M 7.8 on the San Andreas Fault). These are used for emergency drills, public education campaigns (like California's ShakeOut), and loss estimation studies.

Seismic hazard maps are not static products. They require regular updates as new data become available, GMPEs are refined, previously unknown faults are identified, and our understanding of regional tectonics evolves. For example, the U.S. National Seismic Hazard Model is updated roughly every six years to incorporate the latest science.

11.2 Volcanic hazards and monitoring

Volcanic Hazards and Impacts

Types of Volcanic Hazards

Lava flows are streams of molten rock that destroy infrastructure and reshape landscapes. They generally move slowly enough (typically 1–10 km/h for basaltic flows) that people can evacuate, though they leave total destruction in their path.

Pyroclastic flows are the deadliest volcanic hazard. These ground-hugging avalanches of superheated gas, ash, pumice, and rock fragments can exceed 700°C and travel at speeds over 100 km/h. At those temperatures and speeds, survival in the flow path is essentially impossible.

Lahars are volcanic mudflows composed of ash, rock, and water. They form when eruptions melt glacial ice, when heavy rain mobilizes loose volcanic debris, or when crater lakes breach. Lahars can travel tens of kilometers down river valleys, destroying bridges, roads, and buildings far from the volcano itself.

Volcanic ash consists of pulverized rock, minerals, and glass fragments created during explosive eruptions. Its effects are wide-ranging:

- Respiratory problems in humans and animals
- Damage to machinery and electronics (ash is abite and conductive when wet)
- Disruption of aviation by damaging jet engines and reducing visibility (the 2010 Eyjafjallajökull eruption grounded flights across Europe for weeks)
- Roof collapse, especially when ash becomes saturated with rain
- Formation of acid rain when ash mixes with moisture, harming vegetation, acidifying water sources, and corroding metal structures

Impacts on Society and the Environment

Volcanic gases include water vapor, CO_2, SO_2, and H_2S. These can be directly toxic at high concentrations, cause acid rain and air pollution, and influence global climate. Mount Pinatubo's 1991 eruption injected roughly 20 million tonnes of SO_2 into the stratosphere, forming sulfate aerosols that cooled global temperatures by about 0.5°C for 1–2 years.

The human toll depends heavily on hazard type. Slow-moving lava flows (like Kilauea's 2018 eruption) may allow time for evacuation but still destroy property. Pyroclastic flows and lahars are far more lethal because of their speed and intensity.

Environmental effects extend well beyond the eruption site. Ash and gas can alter regional climate patterns, disrupt ecosystems, and contaminate water supplies over large areas. Economic impacts ripple outward too: the Eyjafjallajökull eruption caused an estimated $1.7 billion in losses to the airline industry alone, and eruptions routinely disrupt agriculture, tourism, and global supply chains.

Volcano Monitoring Techniques

Seismic and Deformation Monitoring

Seismicity monitoring uses networks of seismometers to detect earthquakes generated by magma movement beneath a volcano. Volcanic earthquakes tend to have distinct signatures compared to tectonic earthquakes: they're often high-frequency, low-magnitude events clustered in swarms. A sudden increase in seismic swarms, particularly at shallowing depths, is one of the strongest indicators that magma is rising toward the surface.

Ground deformation monitoring tracks physical changes in a volcano's shape. As magma accumulates in a shallow chamber, the overlying surface inflates (bulges outward or tilts). After an eruption, the surface may deflate as magma is withdrawn. Three key tools are used:

- Tiltmeters measure small changes in the slope angle of the volcano's flanks
- GPS stations track precise three-dimensional position changes at the surface
- InSAR (Interferometric Synthetic Aperture Radar) uses repeated satellite radar passes to detect millimeter-scale surface displacement over broad areas

Gas Emission and Remote Sensing Monitoring

Gas emission monitoring measures the composition and flux of gases escaping from a volcano, which provides clues about the depth, volume, and movement of magma. A sharp increase in SO_2 emissions is particularly significant because sulfur tends to exsolve from magma only at shallow depths, signaling that magma is nearing the surface.

Common instruments include:

- UV spectrometers (COSPEC/DOAS) for measuring SO_2 flux
- FTIR spectrometers for detecting CO_2, H_2S, and other gases

Remote sensing supplements ground-based monitoring with broader spatial coverage. Satellite imagery and thermal cameras can detect thermal anomalies (indicating new lava flows or rising heat flux), track ash and gas plumes, and measure surface elevation changes via radar.

The real power of modern volcano monitoring comes from data integration. No single technique is reliable on its own. Scientists combine seismic, deformation, gas, and remote sensing data to build a more complete picture of what's happening beneath a volcano and to assess eruption probability.

Volcanic Hazard Assessment

Hazard Mapping and Risk Communication

Hazard maps delineate areas that could be affected by specific volcanic hazards based on a volcano's eruptive history, surrounding terrain, and numerical modeling. Different hazard types (lava flows, ashfall, lahars, pyroclastic flows) are mapped separately or as overlapping zones, typically color-coded with red indicating the highest risk.

These maps have real limitations. Eruption size, style, and duration carry inherent uncertainty. Topography can change between eruptions (from erosion, construction, or previous deposits), and models may not capture every possible scenario. Hazard maps are best understood as probabilistic guides, not guarantees.

Risk communication translates hazard information into actionable guidance for affected populations through alert levels, warnings, evacuation orders, and public education. Effective communication is:

- Timely so people can act before hazards arrive
- Clear and consistent to avoid confusion
- Delivered by trusted sources such as national volcano observatories and emergency management agencies
- Tailored to local context, accounting for language, cultural norms, and available technology

Challenges and Strategies for Effective Hazard Management

Maintaining public awareness between eruptions is one of the hardest problems in volcanic risk management. Volcanoes with long repose periods can lull nearby communities into complacency. Other persistent challenges include combating misinformation, reaching remote or underserved populations, and balancing safety measures against economic pressures.

Effective hazard management requires collaboration between scientists, government agencies, and local communities. Strategies include:

- Evacuation planning with clearly defined trigger thresholds and routes
- Land-use policies that discourage development in high-risk zones
- Engineering solutions such as lahar diversion barriers and reinforced structures
- Sustained investment in monitoring networks and early warning systems
- Ongoing community engagement through drills, education programs, and regular updates to hazard maps

Volcanic Eruption Case Studies

Mount Pinatubo, Philippines (1991): Successful Hazard Management

The 1991 eruption of Mount Pinatubo was the second-largest eruption of the 20th century. Coordinated monitoring by the Philippine Institute of Volcanology and Seismology (PHIVOLCS) and the U.S. Geological Survey (USGS) detected escalating seismicity and SO_2 emissions weeks before the climactic eruption. Tens of thousands of people were evacuated in time. The advance warning is estimated to have saved over 5,000 lives. Even so, ashfall and subsequent lahars caused massive infrastructure damage, and the eruption's aerosol cloud affected global climate for years.

Nevado del Ruiz, Colombia (1985): Failures in Risk Communication

A relatively small eruption of Nevado del Ruiz melted summit glaciers, generating lahars that traveled over 60 km down river valleys. The town of Armero was buried, killing more than 23,000 people. Scientists had identified the lahar risk, and a preliminary hazard map existed, but critical breakdowns occurred: warnings were delayed, communication between scientists and local authorities was poor, and evacuation orders either weren't issued or weren't received in time. This disaster became a defining case study in why monitoring alone is insufficient without effective communication and emergency response systems.

Eyjafjallajökull, Iceland (2010): Aviation and Economic Disruption

This modest-sized eruption produced a fine ash plume that drifted across European airspace for weeks, grounding over 100,000 flights and stranding roughly 10 million passengers. Total economic losses reached billions of dollars. The event exposed gaps in ash dispersal modeling and forced a rethinking of aviation safety protocols. It also highlighted the tension between precautionary airspace closures and the enormous economic costs of prolonged disruption.

Kilauea, Hawaii (2018): Managing a Long-Duration Eruption

The 2018 lower East Rift Zone eruption of Kilauea destroyed over 700 homes and displaced thousands of residents over several months. Because lava flows moved relatively slowly, evacuations were generally successful. However, the prolonged nature of the eruption created unique challenges: maintaining consistent hazard communication over months, managing "lava tourists" entering danger zones, supporting community resilience, and addressing the mental health impacts of extended displacement. Continuous monitoring by the USGS Hawaiian Volcano Observatory was critical throughout.

11.3 Landslide and slope stability analysis

Slope Stability Factors and Landslide Triggers

Landslides pose significant risks to lives and infrastructure. Slope stability depends on the balance between forces trying to move material downhill and forces resisting that movement. Understanding what controls that balance, and what disrupts it, is the foundation of landslide hazard assessment.

Geophysical methods and remote sensing give us tools to peer beneath the surface and monitor slopes over time. Seismic refraction, electrical resistivity surveys, InSAR, and GIS-based modeling all contribute to identifying vulnerable areas before failure occurs.

Geological and Geomorphological Factors

Slope stability is governed by the interplay of geological, geomorphological, and hydrological conditions. The key material property is shear strength, which depends on two components: cohesion (the "glue" holding particles together) and internal friction angle (resistance to sliding between grains). A slope remains stable as long as shear strength exceeds the shear stress imposed by gravity.

Several geological factors increase landslide susceptibility:

- Weak layers such as clay seams, weathered horizons, or bedding planes oriented parallel to the slope can act as potential slip surfaces
- Rock and soil properties like grain size, mineralogy, and degree of weathering control how strong or weak the slope material is
- Slope geometry matters directly: steeper angles, greater heights, and concave profiles all increase the ratio of driving forces to resisting forces
- Groundwater conditions affect pore water pressure, which reduces effective stress and therefore shear strength

Triggering Mechanisms and External Forces

A slope can sit near the edge of stability for years until a trigger pushes it past the threshold. Common triggers include:

- Intense or prolonged rainfall, which raises pore water pressure and saturates slope materials
- Earthquakes, where seismic ground motion adds transient shear stress and can liquefy saturated soils
- Volcanic activity, including eruptions, lahars, and hydrothermal alteration of rock
- Human-induced changes such as excavation at the toe of a slope, loading at the crest, or altered drainage patterns
- Reservoir filling, which raises the water table and increases pore pressure in adjacent slopes

The factor of safety (FoS) quantifies how close a slope is to failure:

$$FoS = \frac{\text{Resisting forces (shear strength)}}{\text{Driving forces (shear stress)}}$$

- $FoS > 1$: the slope is stable (resisting forces exceed driving forces)
- $FoS = 1$: the slope is at the point of failure
- $FoS < 1$: the slope is unstable and failure is expected

In engineering practice, a FoS of 1.5 or higher is generally required for long-term slope stability to account for uncertainties in material properties and loading conditions.

Geophysical Methods for Landslide Investigation

Seismic Refraction Surveys

Seismic refraction maps the subsurface velocity structure by measuring how fast seismic waves travel through different layers. Because landslide debris is typically looser and more fractured than intact bedrock, there's a velocity contrast that helps delineate the geometry and thickness of the slide mass.

How the method works:

1. An array of geophones is laid out along a line on the slope surface
2. A seismic source (sledgehammer strike or small explosive charge) generates waves at one end
3. The geophones record the arrival times of waves refracted along layer boundaries
4. Travel-time data are inverted to produce a velocity model of the subsurface

What the results reveal:

- Bedrock depth and the thickness of overlying unconsolidated material
- Low-velocity zones that may indicate higher fracture density, water saturation, or weak materials prone to failure
- Groundwater table position, since saturated materials have higher P-wave velocities (close to 1500 m/s for water-saturated sediments)

Electrical Resistivity Surveys

Electrical resistivity imaging maps how easily current flows through the subsurface. Resistivity is sensitive to lithology, water content, and clay content, making it well suited for identifying conditions associated with landslide-prone zones.

Data collection follows a similar surface-array approach:

1. An array of electrodes is placed along a profile on the slope
2. Current is injected through selected electrode pairs while voltage is measured at others
3. Measurements are repeated for many electrode combinations to build up spatial coverage
4. The apparent resistivity data are inverted to produce a 2D or 3D resistivity model

Key interpretive guidelines:

- Low resistivity zones often correspond to water-saturated or clay-rich materials, both of which reduce shear strength
- Time-lapse monitoring (repeating surveys over weeks or months) can track changes in water content or detect the progressive development of slip surfaces
- Sharp resistivity contrasts may mark the boundary between slide material and intact bedrock

Integrating seismic refraction with electrical resistivity provides complementary information: seismic data constrain mechanical properties and layer geometry, while resistivity data highlight fluid content and clay distribution. Both datasets should be interpreted alongside geological mapping and geotechnical borehole data for reliable landslide characterization.

Remote Sensing and GIS in Landslide Analysis

Remote Sensing Techniques for Landslide Mapping

Remote sensing provides high-resolution spatial data for identifying, mapping, and monitoring landslides, especially over large or inaccessible terrain.

- Stereoscopic aerial photography allows 3D visualization of the ground surface, making it possible to identify landslide features like head scarps, tension cracks, and displaced masses
- High-resolution satellite imagery (e.g., WorldView, Pléiades) can detect surface deformation, changes in vegetation cover, and fresh exposure of bare soil associated with recent slides
- LiDAR (Light Detection and Ranging) generates detailed digital elevation models (DEMs) with sub-meter resolution. Because LiDAR can penetrate vegetation canopy, it reveals subtle topographic signatures of old or slow-moving landslides that are invisible in optical imagery
- InSAR (Interferometric Synthetic Aperture Radar) measures ground deformation by comparing the phase of radar signals from repeat satellite passes. It can detect displacement rates as small as a few millimeters per year, making it a powerful tool for monitoring creeping slopes before catastrophic failure

GIS-Based Landslide Susceptibility Assessment

Geographic Information Systems (GIS) serve as the platform for combining all of these data sources into a spatial analysis framework.

The typical workflow for susceptibility mapping:

1. Build a landslide inventory by compiling historical records, field observations, aerial photo interpretation, and satellite-derived detections
2. Assemble thematic layers representing conditioning factors: slope angle, aspect, lithology, land use/land cover, distance to faults, drainage density, and soil type
3. Apply a statistical or machine learning model (logistic regression, random forest, or neural networks) to relate the spatial distribution of past landslides to the conditioning factors
4. Generate a susceptibility map that classifies the landscape into zones of low, moderate, high, and very high landslide probability
5. Validate the model using a portion of the landslide inventory withheld from training

GIS-based approaches combined with remote sensing enable rapid, cost-effective hazard assessment over regional scales, which is particularly valuable in developing countries or mountainous regions where field access is limited.

Case Studies of Significant Landslides

2014 Oso Landslide, Washington, USA

The Oso landslide on March 22, 2014, killed 43 people and destroyed dozens of homes in Snohomish County. The slide occurred in a region with a well-documented history of slope failures, but the scale and mobility of this event were far greater than anticipated.

- The slope consisted of weak, glacially-derived sediments (glaciolacustrine clays and outwash sands) deposited during Pleistocene glaciation
- Weeks of heavy rainfall saturated the slope, raising pore water pressures and reducing effective stress along a pre-existing weak layer
- Post-failure geophysical investigations using seismic refraction and electrical resistivity identified a low-velocity, high-conductivity layer at the base of the slide mass, consistent with a saturated clay-rich slip surface
- The slide mass traveled over 1 km across the valley floor, a runout distance that highlighted the danger of flow-like behavior in saturated, fine-grained debris

2017 Xinmo Landslide, Sichuan, China

On June 24, 2017, a massive rock avalanche buried Xinmo village in Maoxian County, killing over 100 people. This event demonstrated how seismic preconditioning and topographic effects combine to produce catastrophic failures.

- The slope had been weakened by the 2008 Wenchuan earthquake (M_w 7.9) and subsequent aftershocks, which fractured the rock mass extensively
- A smaller, more recent earthquake triggered the final collapse of the heavily fractured ridge
- Seismic monitoring data suggested that topographic amplification of seismic waves along the narrow ridge concentrated shaking at the crest, accelerating the failure
- The resulting rock avalanche had a volume of approximately 13 million cubic meters and buried the village under tens of meters of debris

1963 Vajont Landslide, Italy

The Vajont disaster on October 9, 1963, is one of the most studied landslides in geotechnical history. Approximately 270 million cubic meters of rock slid into the Vajont reservoir, generating a wave that overtopped the dam by over 200 meters and killed nearly 2,000 people downstream.

- The slope above the reservoir contained a pre-existing slow-moving landslide along a clay-rich layer that acted as the basal sliding surface
- As the reservoir was filled, rising water levels increased pore water pressure along this clay layer, reducing effective stress and progressively destabilizing the slope
- Monitoring instruments, including inclinometers in boreholes, recorded accelerating creep in the weeks before failure, but warnings were not acted upon in time
- The dam itself survived intact, but the displaced water devastated the town of Longarone in the valley below

These case studies reinforce a consistent lesson: effective landslide hazard assessment requires a multi-disciplinary approach that integrates geological mapping, geotechnical testing, geophysical surveys, and remote sensing. No single method captures the full picture. The most dangerous situations arise when known risk factors are present but not adequately investigated or communicated.

11.4 Tsunami and coastal hazard assessment

Tsunami Generation and Propagation

Tsunamis are powerful ocean waves triggered by sudden, large-scale displacements of water. They can cross entire ocean basins at jet-aircraft speeds, then pile up into devastating walls of water as they reach the coast. Assessing where, when, and how badly they'll strike is one of the most consequential problems in geophysical hazard analysis.

This section covers the physics of tsunami generation and propagation, methods for reconstructing past events, numerical modeling techniques for hazard mapping, and the design of early warning systems and community preparedness strategies.

Mechanisms of Tsunami Generation

Most tsunamis originate from large, sudden vertical displacements of the seafloor. The three primary generation mechanisms are:

- Subduction zone earthquakes — The most common cause. When the overriding plate snaps upward during fault rupture, it displaces the entire water column above it. The 2011 Tohoku earthquake (M_w 9.1) displaced the seafloor by up to ~5 m over a rupture area roughly 500 km × 200 km.
- Submarine or coastal landslides — A large mass of sediment or rock slides into or beneath the ocean, displacing water. These can generate locally extreme wave heights (the 1958 Lituya Bay landslide produced a run-up of 524 m in the confined bay) but tend to attenuate more rapidly than earthquake-generated tsunamis.
- Volcanic eruptions — Underwater explosions, caldera collapses, or flank collapses can displace water suddenly. The 2022 Hunga Tonga-Hunga Ha'apai eruption generated a tsunami detected across the Pacific.

The initial wave height scales with the amount of vertical seafloor displacement, while the wavelength is controlled by the spatial extent of the source area. A larger rupture area produces a longer-wavelength wave that carries more energy across the open ocean.

Other factors that influence generation efficiency include:

- Water depth at the source (shallower water means less volume to displace, but the coupling between seafloor and surface is more direct)
- How efficiently energy transfers from the solid earth to the water column
- Source directivity, which controls how wave energy is distributed azimuthally

Characteristics of Tsunami Propagation

Tsunamis behave as shallow-water waves because their wavelengths (often 100–500 km in the open ocean) are far greater than the ocean depth (~4 km average). This means the entire water column moves, and the wave speed is governed by:

$$c = \sqrt{g \cdot h}$$

where g is gravitational acceleration and h is water depth. In 4 km of open ocean, this gives $c \approx 200$ m/s (~720 km/h), comparable to a commercial jet. Despite these speeds, open-ocean tsunami amplitudes are typically less than 1 m, making them nearly undetectable at sea.

As a tsunami enters shallower coastal water, wave shoaling occurs:

1. The wave slows down as h decreases.
2. Conservation of energy flux forces the wave amplitude to increase.
3. The wavelength compresses, concentrating energy into a shorter, taller wave.

Several bathymetric effects shape how energy arrives at the coast:

- Refraction — Wave fronts bend toward shallower regions, focusing energy on headlands and spreading it across bays.
- Diffraction — Wave energy spreads around islands and into shadowed areas behind obstacles.
- Resonance — Bays, harbors, and continental shelves can trap and amplify wave energy at their natural oscillation periods. Crescent City, California, is notoriously vulnerable to this effect; during the 1964 Alaska tsunami, harbor resonance amplified waves well beyond what the open-coast signal would suggest.

Paleotsunamis and Historical Records for Hazard Assessment

Because large tsunamis are infrequent at any given location, the instrumental record (roughly the last century) is far too short to capture the full range of possible events. Extending the record backward through paleotsunami research and historical documentation is essential for estimating recurrence intervals and worst-case scenarios.

Paleotsunami Evidence and Analysis Techniques

Paleotsunami deposits are the physical traces left behind when a tsunami floods a coastal area. Common deposit types include:

- Sand sheets — Thin layers of marine sand found inland within fine-grained coastal sediments (marshes, lagoons). Their landward extent provides a minimum estimate of inundation distance.
- Boulder accumulations — Large coral or rock boulders transported onshore by extreme wave energy.
- Erosional unconformities — Surfaces where pre-existing sediment was stripped away by high-velocity flow.

Identifying these deposits and distinguishing them from storm deposits relies on several techniques:

- Sedimentological analysis — Grain size distributions, sorting characteristics, and internal sedimentary structures (e.g., normal grading, rip-up clasts) that are diagnostic of tsunami flow.
- Geochemical markers — Elevated salinity, marine-sourced trace elements, or other saltwater indicators within otherwise freshwater sediments.
- Microfossil assemblages — The presence of marine diatoms, foraminifera, or other organisms in terrestrial or brackish sediment sequences confirms a marine incursion.

Dating methods pin these deposits in time:

- Radiocarbon (^{14}C) dating of organic material above and below the deposit brackets the event age.
- Optically stimulated luminescence (OSL) dates the last time quartz or feldspar grains were exposed to sunlight, useful when organic material is absent.

By mapping multiple paleotsunami deposits along a coastline and dating them, researchers can reconstruct recurrence intervals. The Cascadia Subduction Zone, for example, has a paleotsunami record showing roughly 19 major events over the past 10,000 years, yielding an average recurrence interval of ~500 years.

Integration of Historical Records and Modern Modeling

Historical records fill the gap between the paleotsunami record and the instrumental era. Sources include:

- Written eyewitness accounts (the 1755 Lisbon tsunami is documented extensively in European archives)
- Photographs and newspaper reports from more recent events
- Tide gauge records, which provide quantitative wave arrival times and amplitudes (the 1960 Chile tsunami was recorded on tide gauges across the Pacific)

Combining paleotsunami data, historical records, and modern source characterization enables probabilistic tsunami hazard assessment (PTHA). PTHA estimates the probability of exceeding a given wave height at a specific location over a defined time period, analogous to probabilistic seismic hazard analysis. The outputs are hazard curves and maps that inform risk management decisions, land-use planning, and building code requirements.

Numerical Modeling for Tsunami Inundation and Hazard Mapping

Numerical Models and Governing Equations

Numerical models simulate the full lifecycle of a tsunami: generation, open-ocean propagation, and coastal inundation. The workflow generally follows these steps:

1. Define the source — Specify earthquake fault parameters (location, geometry, slip distribution) or landslide volume and geometry.
2. Compute initial sea-surface displacement — For earthquakes, this is typically done using elastic dislocation models (e.g., Okada, 1985).
3. Propagate the wave — Solve the governing equations on a computational grid that covers the ocean basin.
4. Resolve coastal inundation — Use high-resolution grids near the coast to capture shoaling, run-up, and flooding.

The most commonly used governing equations are:

- Nonlinear shallow-water equations (NLSWE) — These capture the essential physics of long-wave propagation and are computationally efficient. Most operational tsunami models (e.g., MOST, ComMIT, GeoClaw) are based on NLSWE.
- Boussinesq equations — These add dispersive terms that become important for shorter-wavelength tsunamis (e.g., landslide-generated waves) where frequency dispersion affects wave shape during propagation.

Key model inputs include high-resolution bathymetry (seafloor topography) and coastal topography (elevation data for inundation modeling).

Model Outputs and Applications

Model outputs include:

- Maximum wave heights at each grid point
- Flow velocities and inundation depths on land
- Arrival times at coastal locations

These outputs feed directly into hazard mapping and risk assessment:

- Nested grid approaches balance computational cost and resolution. A coarse grid covers the open ocean (perhaps 1–4 km resolution), intermediate grids cover the regional shelf (~100–500 m), and fine grids resolve harbors and coastal communities (~10–30 m).
- Hazard maps delineate zones of expected inundation for scenario events or probabilistic return periods. These maps guide evacuation zone boundaries, land-use planning, building codes, and insurance rates.

Model validation is critical. Modelers benchmark their codes against:

- Analytical solutions for idealized problems
- Laboratory wave-tank experiments
- Observed data from historical tsunamis (tide gauge records, field survey measurements of run-up and inundation)

Without rigorous validation, hazard maps derived from models cannot be trusted for life-safety decisions.

Tsunami Early Warning Systems and Community Preparedness

Components and Effectiveness of Early Warning Systems

Tsunami early warning systems aim to detect a tsunami-generating event, estimate the threat, and deliver actionable warnings before the waves arrive. The basic chain of operations is:

1. Seismic detection — Seismometers detect the earthquake and rapidly estimate its location, depth, and magnitude. For tsunamigenesis, the key parameters are whether the earthquake is shallow, submarine, and large enough (generally $M_w \geq 7.0$ for significant tsunamis).
2. Sea-level monitoring — Deep-ocean pressure sensors (DART buoys, maintained by NOAA and partner agencies) detect the passage of the tsunami wave in the open ocean. Coastal tide gauges confirm wave arrivals and amplitudes.
3. Threat assessment — Pre-computed model databases are queried in real time. The observed DART data constrain which scenario best matches the actual event, allowing rapid refinement of wave height forecasts.
4. Warning dissemination — Alerts are issued through national warning centers (e.g., PTWC, JMA) to emergency management agencies, which activate sirens, broadcast alerts, and initiate evacuations.

The effectiveness of the system depends on:

- Station density and distribution — Gaps in the monitoring network create blind spots.

• Processing speed — For near-field tsunamis (source close to the coast), warning times may be only minutes. The 2018 Sulawesi tsunami struck Palu within ~10 minutes of the earthquake, leaving almost no time for a formal warning.

• Communication reliability — Warnings must reach the public through multiple redundant channels (sirens, cell alerts, radio, TV).

• False alarm management — Too many false alarms erode public trust and reduce compliance. Too few warnings risk missing real events. Robust detection algorithms and clear communication protocols are essential.

Community Preparedness and Mitigation Strategies

Even the best warning system is useless if communities don't know how to respond. Preparedness has several components:

• Public education — Residents and visitors in tsunami-prone areas need to know the natural warning signs (strong earthquake shaking, unusual ocean withdrawal) and the correct response (move to high ground immediately without waiting for an official alert).

• Evacuation planning — Hazard maps define evacuation zones. Evacuation routes should be clearly signed, lead to high ground or designated assembly areas, and be designed to handle the expected population flow.

• Regular drills — Practicing evacuations maintains awareness and reveals logistical problems before a real event. Japan conducts annual tsunami drills on the anniversary of past disasters, and participation rates directly correlate with evacuation compliance during actual events.

• Vertical evacuation structures — In flat coastal areas where high ground is too far away, reinforced concrete buildings or purpose-built evacuation towers provide refuge above expected inundation levels. These structures must be engineered to withstand both the hydrodynamic forces and debris impact of a tsunami.

• Resilient infrastructure — Seawalls, breakwaters, and coastal forests can reduce (but not eliminate) tsunami energy. The 2011 Tohoku tsunami overtopped many seawalls designed for smaller events, demonstrating that structural defenses alone are insufficient.

Post-event assessment after every significant tsunami provides feedback for improving the entire system. Field surveys measure actual run-up and inundation, which are compared against model predictions. Interviews with survivors reveal how warnings were received and how evacuations unfolded. Lessons from the 2011 Tohoku tsunami, for example, led to major revisions in Japan's hazard maps, seawall design standards, and evacuation protocols.

Unit 12 – Geodynamics and Earth's Interior

12.1 Earth's internal structure and composition

Earth's interior is a layered system with distinct physical and chemical properties at each depth. Understanding this structure is foundational for geodynamics: it explains why plates move, how Earth generates its magnetic field, and what drives volcanic activity and seismicity.

Geophysicists probe the interior using seismic waves, gravity measurements, magnetic surveys, and heat flow analysis. These methods reveal a planet that is chemically stratified (crust, mantle, core) and mechanically stratified (lithosphere, asthenosphere, mesosphere), with pressure and temperature jointly controlling the behavior of materials at every depth.

Earth's Internal Structure

Major Layers and Their Characteristics

Earth's interior divides into three chemically distinct layers: the crust, mantle, and core.

- Crust: The outermost layer, 5–70 km thick, composed of relatively low-density silicate rocks (granite in continental crust, basalt in oceanic crust).
- Mantle: Extends from the base of the crust to ~2,900 km depth. Composed of hot, dense ultramafic rock, predominantly peridotite.
- Core: Extends from ~2,900 km to Earth's center at ~6,371 km. Composed of iron-nickel alloy. The outer core is liquid, while the inner core is solid. The inner core remains solid despite higher temperatures because the immense pressure (over 300 GPa) raises the melting point of iron above the local temperature.

Lithosphere and Asthenosphere

The chemical layering above doesn't fully capture how the Earth behaves mechanically. For that, you need the rheological layering.

- The lithosphere includes the crust and the uppermost mantle. It behaves as a rigid, brittle shell, typically 70–150 km thick (thinner beneath ocean ridges, thicker beneath old continental interiors).
- The asthenosphere lies beneath the lithosphere, extending to roughly 200–300 km depth. It's solid rock, but at temperatures close to its melting point, so it deforms in a ductile, viscous manner over geological timescales.
- The boundary between them is defined by a change in rheology (mechanical behavior), not composition. This transition is what makes plate tectonics possible: rigid lithospheric plates can slide over the weaker asthenosphere.

Methods for Studying Earth's Interior

Seismic Waves

Seismic waves are the primary tool for imaging Earth's interior. They're generated by earthquakes or controlled-source explosions and change speed, direction, and amplitude as they pass through materials with different elastic properties and densities.

Two key body-wave types:

- P-waves (primary/compressional): Travel through both solids and liquids. They arrive first because they're faster (~6 km/s in the upper mantle, ~13 km/s near the base of the mantle).
- S-waves (secondary/shear): Travel only through solids. The fact that S-waves do not propagate through the outer core is the direct evidence that it's liquid.

How boundaries are detected:

- Seismic waves reflect and refract at interfaces where material properties change abruptly. The Mohorovičić discontinuity (Moho) marks the crust-mantle boundary, identified by a sharp increase in P-wave velocity. The core-mantle boundary (CMB) at ~2,900 km produces strong reflections and the S-wave shadow zone.
- Seismic tomography inverts travel-time data from many earthquakes recorded at many stations to build 3D velocity models of the interior. Regions with anomalously fast velocities typically correspond to colder, denser material (e.g., subducting slabs), while slow anomalies suggest hotter, less dense material (e.g., mantle plumes).

Other Geophysical Methods

- Gravity measurements (from satellites like GRACE/GOCE and surface gravimeters) map lateral density variations. Positive gravity anomalies indicate denser subsurface material; negative anomalies suggest lower-density regions.
- Magnetic surveys detect variations in Earth's magnetic field. At the surface, these reflect the distribution of magnetic minerals in crustal rocks. At a global scale, the field's geometry and secular variation constrain models of fluid flow in the electrically conductive outer core.
- Heat flow measurements quantify the rate of heat loss through Earth's surface (global average ~87 mW/m^2). These data constrain the thermal structure of the lithosphere and the vigor of mantle convection beneath it.

Composition and Properties of Earth's Layers

Crust

The crust is composed primarily of silicate minerals, but oceanic and continental crust differ significantly.

- Oceanic crust: 5–10 km thick, mafic in composition (rich in Mg and Fe), dominated by basalt and gabbro. It forms at mid-ocean ridges by partial melting of the underlying mantle. Average density is ~3.0 g/cm^3.
- Continental crust: 30–70 km thick, felsic in composition (rich in Si and Al), with an average composition close to granodiorite. It forms through a complex history of magmatic accretion, metamorphism, and differentiation over billions of years. Average density is ~2.7 g/cm^3.

This density contrast is why continental crust "floats" higher on the mantle (isostasy) and is not easily subducted.

Mantle

The mantle makes up ~84% of Earth's volume. Its dominant rock type is peridotite, rich in olivine and pyroxene.

- The upper mantle (down to ~410 km) is relatively cooler and more rigid. Xenoliths brought up by volcanic eruptions provide direct samples of this region.
- The transition zone (410–660 km) is defined by pressure-induced mineral phase changes: olivine transforms to wadsleyite at ~410 km and then to ringwoodite at ~520 km. These transitions produce sharp increases in seismic velocity and density.
- The lower mantle (660–2,900 km) is hotter and composed of high-pressure phases, primarily bridgmanite (Mg-silicate perovskite structure) and ferropericlase. Near the CMB, bridgmanite may further transform to post-perovskite, which influences the dynamics of the lowermost mantle (D" layer).

Mantle convection, driven by heat from the core and internal radioactive decay (primarily U, Th, K), is the engine of plate tectonics. Convective flow transports heat from the interior to the surface and drives the motion of lithospheric plates. Partial melting of the mantle at ridges and above subduction zones generates the magmas that feed volcanism and build new crust.

Core

The core is composed primarily of iron-nickel alloy, with a small fraction of lighter elements (likely S, O, Si, or some combination). Evidence for its composition comes from seismic velocities, Earth's bulk density (~5.5 g/cm³ vs. crustal rocks at ~2.7–3.0 g/cm³), and comparison with iron meteorites.

- Outer core (~2,900–5,150 km): Liquid. Its density ranges from ~9.9 to ~12.2 g/cm³. Convective motions of this electrically conductive fluid generate Earth's magnetic field through a self-sustaining geodynamo.
- Inner core (~5,150–6,371 km): Solid, with a density of ~13 g/cm³. It's slowly growing as the Earth cools and the outer core crystallizes at the inner core boundary (ICB). This crystallization releases latent heat and light elements into the outer core, helping to drive the convection that sustains the geodynamo.

Pressure and Temperature's Influence on Earth's Interior

Pressure Effects

Pressure increases with depth due to the weight of overlying material. At the center of the Earth, pressure reaches approximately 360 GPa (about 3.6 million atmospheres).

- Increasing pressure raises the density and seismic velocity of materials, even without a change in composition.
- Pressure drives mineral phase transitions that define key boundaries in the mantle. For example, olivine (α-phase) transforms to wadsleyite (β-phase) near 410 km depth, and ringwoodite (γ-phase) transforms to bridgmanite + ferropericlase near 660 km. Each transition involves a denser crystal structure accommodating the same bulk chemistry.
- In the lower mantle, high-pressure phases like bridgmanite and post-perovskite have distinct elastic properties that influence seismic wave propagation and mantle dynamics.

Temperature Effects

Temperature increases with depth, but the geothermal gradient is not constant.

- In the lithosphere, the gradient is steep (~25–30 °C/km near the surface).
- In the convecting mantle, the gradient is much gentler (close to the adiabatic gradient, ~0.3–0.5 °C/km), because convection efficiently homogenizes temperature.
- Sharp thermal boundary layers exist at the base of the lithosphere and at the CMB, where temperature jumps by an estimated 1,000–1,500 K over a relatively thin zone.

High temperatures reduce viscosity and promote ductile flow, enabling mantle convection. They also control where partial melting occurs: when the local temperature exceeds the solidus of mantle rock (which itself depends on pressure and composition), melt is generated.

Combined Pressure and Temperature Effects

Pressure and temperature act together to determine the phase, rheology, and density of materials at any given depth.

- The inner core is solid not because it's cold, but because pressure raises the melting point of iron above the local temperature. The outer core is liquid because pressure is lower there, and the temperature exceeds the melting point.
- Mantle convection patterns depend on the pressure- and temperature-dependent viscosity of mantle rock. Viscosity variations of several orders of magnitude exist between the upper and lower mantle.
- Unique mineral assemblages stable only under extreme P-T conditions (e.g., post-perovskite near the CMB) influence seismic anisotropy, heat transport, and the dynamics of thermal boundary layers.

Understanding how pressure and temperature jointly control material behavior is essential for modeling mantle convection, the geodynamo, and the long-term thermal evolution of the planet.

12.2 Mantle convection and plate tectonics

Mantle Convection and Driving Forces

Thermal Convection and Gravitational Instability

Mantle convection is the slow, continuous circulation of Earth's mantle driven by heat transfer from the core to the surface. It's the engine behind plate tectonics, and understanding it is central to explaining why Earth's surface looks and behaves the way it does.

Two forces drive this circulation:

- Thermal convection results from temperature differences between the hot lower mantle and the cooler upper mantle. Hotter material is less dense, so it rises; cooler material is denser, so it sinks. This sets up a cyclical flow.
- Gravitational instability reinforces this pattern. Wherever density contrasts exist, gravity pulls denser material downward and allows buoyant material to ascend. Cold, dense subducting slabs are a prime example of gravitational instability at work.

Convective flow velocities in the mantle are slow on human timescales, typically a few centimeters per year (comparable to how fast your fingernails grow), though some estimates reach several tens of centimeters per year in regions of vigorous flow.

Heat Generation and Convective Patterns

The mantle's heat budget has two main sources: primordial heat left over from Earth's formation, and ongoing radioactive decay of uranium, thorium, and potassium within the mantle. This internal heat production helps sustain convection over billions of years.

Convective flow organizes into recognizable patterns:

- Large-scale convection cells involve broad regions of rising hot material and sinking cool material, analogous to circulation in a pot of heated fluid, but operating over thousands of kilometers.
- Mantle plumes are narrower, localized columns of anomalously hot material that originate near the core-mantle boundary (~2,900 km depth) and rise through the mantle. These are distinct from the broader cell-like circulation.

Numerical models that incorporate mantle viscosity, temperature-dependent rheology, and internal heating reproduce patterns that closely match observed seismic tomography images and geoid anomalies, giving us confidence that our understanding of mantle dynamics is on the right track.

Mantle Convection and Plate Tectonics

Driving Mechanism and Plate Boundaries

Mantle convection provides the forces that move lithospheric plates. The convective flow exerts drag on the base of plates, and density contrasts within the plates themselves (particularly at subduction zones) contribute additional driving forces like slab pull and ridge push.

The three types of plate boundaries correspond to different parts of the convective system:

- Divergent boundaries form where mantle material rises and spreads laterally, pulling plates apart. New oceanic crust is created at mid-ocean ridges (e.g., the East Pacific Rise). Ridge push, caused by the elevated topography and warm, buoyant material at ridges, helps drive plates away from these boundaries.
- Convergent boundaries form where cooler, denser lithosphere sinks back into the mantle. This produces subduction zones, oceanic trenches, and volcanic arcs (e.g., the Andes Mountains). Slab pull, the gravitational force on the dense descending slab, is thought to be the single largest force driving plate motion.
- Transform boundaries develop where plates slide horizontally past each other, accommodating differential motion between adjacent plates. The San Andreas Fault is a classic example.

Plate Boundary Patterns and Mantle Convection Cells

The global distribution of plate boundaries aligns well with the predicted geometry of mantle convection. Subduction zones and oceanic trenches sit above descending limbs of convective flow, while mid-ocean ridges overlie ascending regions.

The geometry and orientation of plate boundaries are influenced by the underlying mantle flow, but the relationship runs both ways. Subducting slabs cool the mantle locally and drive downwelling, while ridges passively respond to plate separation rather than being pushed apart by rising plumes in most cases. This creates a coupled, self-sustaining system: convection moves plates, and plate motions (especially subduction) help organize convection.

Evidence for Mantle Convection

Seismic Tomography

Seismic tomography images Earth's interior by measuring how fast seismic waves travel through different regions of the mantle. Variations in wave speed reflect differences in temperature, composition, and density.

Here's how to interpret the results:

1. Seismic waves travel faster through cold, dense material and slower through hot, less dense material.

2. High-velocity anomalies (shown as blue in tomographic images) indicate colder, denser regions, typically corresponding to subducting slabs.

3. Low-velocity anomalies (shown as red) indicate hotter, less dense regions, corresponding to upwelling zones or mantle plumes.

4. By compiling data from thousands of earthquakes recorded at stations worldwide, tomographic models build 3D maps of these velocity variations throughout the mantle.

These images reveal large-scale structures consistent with convection: descending high-velocity slabs beneath subduction zones and low-velocity columns extending from near the core-mantle boundary toward the surface beneath known hot spots.

Geoid Anomalies

The geoid is the shape Earth's surface would take if covered entirely by ocean, determined by the gravitational field. Variations in the geoid reflect density differences within Earth's interior.

- Positive geoid anomalies (geoid surface bulges outward) generally correspond to regions of higher-than-average density at depth, such as subducting slabs, though dynamic effects from mantle flow also contribute.

- Negative geoid anomalies (geoid surface dips inward) tend to correspond to lower-density regions.

The long-wavelength geoid shows a dominant degree-2 pattern, meaning two broad highs and two broad lows around the globe. This pattern is consistent with large-scale mantle convection and correlates with the distribution of major upwelling and downwelling regions. The match between geoid observations and convection models provides independent support for mantle circulation beyond what seismic tomography alone reveals.

Mantle Plumes and Hot Spot Volcanism

Characteristics and Examples

Mantle plumes are localized upwelling columns of anomalously hot material that originate near the core-mantle boundary and rise buoyantly through the mantle. When a plume reaches the base of the lithosphere, it can cause partial melting, producing volcanic activity at the surface.

This hot spot volcanism is distinct from volcanism at plate boundaries because it occurs in plate interiors (or at boundaries only by coincidence). Key examples include:

- Hawaiian Islands: The classic hot spot chain, sitting in the middle of the Pacific Plate, far from any plate boundary.

• Iceland: Located on the Mid-Atlantic Ridge, where a mantle plume coincides with a divergent boundary, producing unusually voluminous volcanism.

• Yellowstone Caldera: A continental hot spot responsible for massive explosive eruptions and ongoing geothermal activity.

Hot spot lavas have distinct geochemical signatures compared to mid-ocean ridge basalts. They tend to be enriched in incompatible elements (elements that don't fit easily into common mantle minerals), pointing to a deep mantle source that has remained relatively isolated from the convective mixing that homogenizes the upper mantle.

Implications for Mantle Dynamics and Plate Motion

Hot spots are approximately fixed relative to the deep mantle, while lithospheric plates move over them. This means a single plume can produce a chain of progressively older volcanoes stretching across a plate.

The Hawaiian-Emperor Seamount Chain is the best example. The active volcano (Kilauea) sits over the plume today, while islands and seamounts become progressively older to the northwest, reaching ages of ~80 million years at the far end of the Emperor chain. The ~60° bend in the chain at about 47 Ma records a major change in Pacific Plate motion direction.

By measuring the ages and positions of volcanoes along hot spot tracks, you can reconstruct absolute plate velocities, meaning the motion of plates relative to the deep mantle rather than just relative to each other.

The existence of plumes originating near the core-mantle boundary also tells us something fundamental about mantle structure: convection is not confined to the upper mantle alone. Plumes provide evidence for whole-mantle convection, where material circulates between the surface and the deepest parts of the mantle. The buoyancy of rising plumes may also exert forces on the overlying plates, though this effect is generally smaller than slab pull.

12.3 Core dynamics and the geomagnetic field

Earth's Core Structure and Composition

Layers and Boundaries

Earth's core has two distinct layers separated by a sharp boundary. The outer core is a liquid shell roughly 2,260 km thick, sitting beneath the core-mantle boundary (CMB) at about 2,890 km depth. Below it, the inner core is a solid sphere with a radius of about 1,220 km. The boundary between them, the inner core boundary (ICB), is where liquid iron freezes onto the growing inner core.

Composition and Physical Properties

Both layers are composed primarily of iron and nickel, but their physical states differ because of the extreme pressure gradient through the core.

• The outer core is liquid and contains lighter alloying elements (sulfur, oxygen, silicon, and possibly hydrogen). These light elements lower the melting point and play a key role in driving convection.

• The inner core is solid despite being hotter, because the pressure at Earth's center (around 360 GPa) is high enough to force iron into a solid phase. It is compositionally purer iron-nickel than the outer core, since lighter elements are expelled as the inner core crystallizes.

- Core temperatures range from roughly 4,000 K at the CMB to about 5,000–6,000 K at the center. The pressure increases from ~136 GPa at the CMB to ~364 GPa at Earth's center.

The density jump at the ICB (from ~12,200 kg/m³ in the outer core to ~13,000 kg/m³ in the inner core) was first identified through seismic observations by Inge Lehmann in 1936.

Geodynamo and Earth's Magnetic Field

Geodynamo Process

Earth's magnetic field originates in the liquid outer core through the geodynamo process. The basic requirement is simple: you need a rotating body containing a large volume of electrically conducting fluid undergoing vigorous convection.

Convection in the outer core is driven by two main energy sources:

- Thermal buoyancy: Heat released at the ICB (both from the latent heat of inner core crystallization and from secular cooling of the core) creates temperature differences that drive fluid motion.

- Compositional buoyancy: As iron crystallizes onto the inner core, lighter elements (O, S, Si) are rejected into the outer core fluid. This lighter fluid is buoyant and rises, providing a powerful and efficient source of convection.

Earth's rotation organizes these convective flows through the Coriolis effect, producing columnar, helical flow structures aligned roughly parallel to the rotation axis. These are sometimes called Taylor columns.

Dynamo Effect and Magnetic Field Generation

The dynamo effect is a positive feedback loop. Here's how it sustains itself:

1. Convective motions move the electrically conducting iron-alloy fluid through a pre-existing (even very weak) magnetic field.

2. This motion induces electric currents in the fluid (by Faraday's law of electromagnetic induction).

3. Those electric currents generate their own magnetic field, reinforcing and reshaping the original field.

4. The reinforced field interacts with continued fluid motion, sustaining the cycle.

The key parameter governing whether a dynamo can operate is the magnetic Reynolds number, $Re_m = \frac{UL}{\eta}$, where U is a typical flow velocity, L is a characteristic length scale, and η is the magnetic diffusivity. For self-sustaining dynamo action, Re_m must exceed a critical value (typically $Re_m \gtrsim 10\text{–}100$). In Earth's outer core, Re_m is estimated to be on the order of several hundred, well above the threshold.

The geodynamo is inherently a chaotic, nonlinear system. Small changes in flow patterns can produce large changes in field geometry over time, which is why the field fluctuates in strength and direction and occasionally reverses polarity entirely.

Evidence for Core Dynamics

Secular Variation

Secular variation refers to changes in Earth's magnetic field on timescales of years to centuries. You can observe it directly: the position of magnetic north drifts measurably over decades, and the field intensity at any given location changes over time.

- The westward drift of magnetic field features (about 0.2° per year on average) suggests that flow patterns in the outer core are not static but evolve continuously.
- Regional changes in field intensity, such as the growth of the South Atlantic Anomaly (a region of unusually weak field strength), provide a window into how convection patterns in the outer core shift over human timescales.

Secular variation is the most direct evidence that the core is dynamically active right now.

Paleomagnetic Records

On longer timescales, the history of Earth's magnetic field is preserved in rocks and sediments. When igneous rocks cool through the Curie temperature of their magnetic minerals (e.g., magnetite at ~580°C), they lock in the direction and intensity of the ambient field. Sediments can also acquire a magnetic signature as magnetic grains align with the field during deposition.

These paleomagnetic records reveal several important features of core dynamics:

- Polarity reversals: The north and south magnetic poles have swapped positions hundreds of times over Earth's history. The most recent reversal, the Brunhes-Matuyama reversal, occurred about 780,000 years ago.
- Irregular reversal frequency: The average interval between reversals is roughly 200,000–300,000 years, but this varies enormously. During the Cretaceous Normal Superchron (~84–124 Ma), the field maintained a single polarity for about 40 million years. At other times, reversals occurred every few tens of thousands of years.
- Apparent polar wander paths: Because paleomagnetic data record the paleolatitude of a rock at the time it formed, tracking the apparent position of the magnetic pole over time for a given continent reveals how that continent has moved. This was one of the early lines of evidence supporting plate tectonics.

The geomagnetic polarity timescale (GPTS), constructed from marine magnetic anomaly patterns on the seafloor and radiometric dating of volcanic rocks, provides a detailed chronology of reversals extending back roughly 170 million years.

Core Dynamics: Implications for Earth's Evolution

Thermal and Chemical Evolution

The core doesn't exist in isolation. Its thermal and chemical evolution is coupled to the rest of the planet.

- Heat flowing out of the core across the CMB helps drive mantle convection, which in turn drives plate tectonics. The estimated heat flux across the CMB is roughly 5–15 TW, a significant fraction of Earth's total heat budget.
- The inner core is growing over time as the core slowly cools. Current estimates suggest the inner core nucleated somewhere between 0.5 and 1.5 billion years ago, though this remains debated. Before inner core nucleation, the geodynamo would have been powered solely by thermal convection, which is less efficient than the compositional convection available today.
- Changes in core cooling rate, influenced by mantle convection patterns and the thermal insulating effect of large low-shear-velocity provinces (LLSVPs) at the base of the mantle, may explain long-term variations in reversal frequency.

Magnetic Field Reversals and Their Consequences

During a polarity reversal, the dipole field doesn't simply flip instantaneously. The transition takes roughly 1,000–10,000 years, during which the field weakens substantially (perhaps to 10–25% of its normal strength) and becomes dominated by complex, non-dipolar components.

This weakening matters because Earth's magnetic field deflects charged particles from the solar wind and cosmic rays. A significantly weakened field could lead to:

- Increased radiation exposure at Earth's surface, potentially raising mutation rates in organisms
- Changes in atmospheric chemistry, particularly increased production of cosmogenic isotopes like ^{10}Be and ^{14}C (which are actually used to detect past field weakening in ice cores and sediments)
- Possible effects on ozone concentrations, though the magnitude of this effect is still debated

Whether reversals have caused measurable biological extinctions remains an open question. The paleontological record does not show a clear correlation between reversals and mass extinctions, but subtler ecological effects are harder to rule out.

For modern civilization, even a partial weakening of the field (not necessarily a full reversal) poses practical risks to satellite electronics, power grid stability, and navigation systems that rely on the geomagnetic field. Understanding the dynamics of the core is therefore not just an academic exercise but relevant to predicting how the field will behave in coming centuries.

12.4 Rheology and deformation of Earth materials

Rheology of Earth's Interior

Rheology is the study of how materials deform and flow under applied stress. For geodynamics, it's one of the most important concepts because it governs how Earth's interior actually behaves: whether rock flows, bends, or snaps. Rheological properties control mantle convection, plate tectonics, earthquake generation, and the formation of geological structures like folds and faults.

The key insight is that the same rock can behave very differently depending on conditions. A granite near the surface will fracture under stress, but that same granite buried 30 km deep at 600°C will slowly flow. Rheology explains why, and the factors driving these differences (temperature, pressure, composition, strain rate) all change systematically with depth.

Definition and Importance

Rheology describes how materials respond to applied forces over time. For Earth materials like rocks, minerals, and partially molten material, rheological behavior determines whether they deform elastically, flow viscously, or fracture.

This matters because Earth's interior isn't static. The mantle convects, plates move, and faults slip. All of these processes depend on the mechanical response of rock to stress, and that response is what rheology quantifies.

Relevance to Geodynamic Processes

- Rheological properties control the flow and deformation of materials in the mantle and lithosphere
- The viscosity structure of the mantle determines the pattern and vigor of convection

- Rheology influences the rate of plate motion, the localization of deformation at plate boundaries, and the depth at which earthquakes nucleate

Rheological Behavior Types

Elastic and Plastic Deformation

Elastic deformation occurs when a material deforms under stress but returns to its original shape once the stress is removed. Think of it as reversible deformation. Rocks under small stresses and at short timescales typically behave elastically. This is the behavior that stores energy before an earthquake.

Plastic deformation is permanent shape change without fracturing. The material yields beyond a threshold stress (the yield strength) and doesn't recover its original geometry. Deep in Earth's interior, where temperatures and pressures are high, rocks commonly deform plastically rather than breaking.

Viscous, Ductile, and Brittle Deformation

Viscous deformation involves continuous flow under applied stress, where the rate of deformation depends on the viscosity of the material. The mantle behaves as a viscous fluid over geological timescales (millions of years), even though it's solid rock. Mantle viscosity is on the order of 10^{19} to 10^{21} Pa·s, enormously higher than everyday fluids but low enough to permit convective flow over millions of years.

Ductile deformation is the broad term for slow, continuous deformation without fracturing. It dominates in the lower crust and mantle where temperatures and pressures are high. Ductile shear zones are a common expression of this behavior.

Brittle deformation occurs when rock fractures or breaks. This is typical at low temperatures and low confining pressures near Earth's surface (the upper crust). Faults and joints are products of brittle deformation, and brittle failure along faults is what generates earthquakes.

The brittle-ductile transition marks the depth at which deformation style shifts from dominantly brittle to dominantly ductile. In continental crust, this typically occurs around 10–20 km depth, depending on composition and geothermal gradient.

Factors Influencing Rheology

Temperature and Pressure Effects

Temperature is the single most important factor controlling rock rheology.

- Higher temperatures lower viscosity and promote ductile flow. This is because thermal energy activates crystal-scale deformation mechanisms (like dislocation climb) that allow rock to creep.
- The geothermal gradient (the rate of temperature increase with depth, typically ~25–30°C/km in continental crust near the surface) means that deeper rocks are progressively weaker and more prone to flow.

Pressure also plays a major role, though its effects are more nuanced.

- Increasing confining pressure (from the weight of overlying rock) tends to suppress fracturing by pressing crack surfaces together, promoting ductile behavior instead.
- However, pressure also increases the strength of some deformation mechanisms, so its net effect depends on the specific conditions.

- At great depth, the combined effect of high temperature and high pressure determines which deformation mechanism dominates.

Composition and Strain Rate

Composition directly affects rock strength and deformation style.

- Different minerals have very different rheologies. Quartz is relatively weak and begins to flow at moderate temperatures (~300°C), while olivine, the dominant mineral in the upper mantle, is much stronger and requires higher temperatures (~700°C+) to deform ductilely.

- The presence of fluids (water, partial melt) dramatically weakens rocks. Water in crystal lattices reduces the activation energy for creep, and even small amounts of partial melt can reduce bulk viscosity by orders of magnitude.

Strain rate is how fast deformation occurs, and it determines which behavior you observe.

- At low strain rates (slow deformation), rocks have time to flow ductilely. This is the regime relevant to mantle convection and long-term tectonic deformation.

- At high strain rates (rapid deformation), the same rock may behave brittlely because it can't flow fast enough to accommodate the stress. Earthquake rupture is an extreme example of high strain rate deformation.

- Strain rate also controls which creep mechanism dominates: diffusion creep (atom-by-atom migration, dominant at low stress and small grain size) versus dislocation creep (movement of crystal defects, dominant at higher stress). These mechanisms have different dependencies on temperature, grain size, and stress, which is why mantle flow laws are written separately for each.

Rheology in Geodynamic Processes

Mantle Convection and Plate Tectonics

Mantle convection is driven by density differences from thermal gradients, but the pattern and vigor of that convection are controlled by mantle rheology.

- Mantle viscosity varies with depth. The asthenosphere (roughly 100–300 km depth) has relatively low viscosity, partly due to small amounts of partial melt or water. The lower mantle is more viscous, roughly 10–100 times higher than the upper mantle.

- These viscosity contrasts may contribute to partially layered convection, where upper and lower mantle circulate somewhat independently, though whole-mantle flow also occurs.

Plate tectonics depends on a strong lithosphere overlying a weaker asthenosphere.

- The lithosphere is rheologically strong (cool, high viscosity) and behaves as a rigid plate.

- The asthenosphere is rheologically weak (warm, lower viscosity) and allows plates to move over it.

- Rheological weakening at plate boundaries, caused by fluids, elevated temperatures, or grain-size reduction, localizes deformation and enables plate boundaries to exist as narrow zones rather than broad regions of distributed strain.

Earthquakes and Geological Structures

Earthquakes are fundamentally a rheological phenomenon. They occur when rocks in a fault zone undergo a transition from slow, stable sliding (or locking) to rapid brittle failure.

- Fault zones contain materials with specific frictional properties that control whether a fault creeps stably or accumulates elastic strain that releases suddenly.
- Changes in fluid pressure within fault zones can reduce effective normal stress, weakening the fault and potentially triggering slip. This is why fluid injection (e.g., from wastewater disposal) can induce earthquakes.
- Temperature variations along a fault with depth determine where the fault is locked (shallow, brittle) versus where it creeps (deeper, ductile).

Geological structures record the rheological conditions under which they formed.

- Brittle faults and fracture networks form in cool, shallow rock under rapid or high-magnitude stress.
- Ductile shear zones and folds form in warm, deep rock deforming slowly.
- The geometry and style of structures in mountain belts and sedimentary basins reflect the interplay between tectonic stress and the rheological stratification of the crust and upper mantle.

Unit 13 – Geophysical Field Methods and Instrumentation

13.1 Survey design and planning

Survey Design Principles

Geophysical survey design is the process of selecting methods, defining parameters, and planning data acquisition strategies so that field investigations actually answer the questions they're meant to answer. Poor planning leads to gaps in coverage, wasted time, and data that can't be interpreted. A well-designed survey balances scientific goals against practical constraints like budget, terrain, and available equipment.

Selecting Appropriate Methods and Parameters

The starting point is understanding your target: its expected depth, size, geometry, and the physical property contrast that distinguishes it from surrounding material. These factors determine which geophysical methods will work and what resolution you need.

- Choose methods sensitive to the relevant physical property contrast. Seismic reflection works well for imaging deep sedimentary layers because of acoustic impedance contrasts, while ground-penetrating radar (GPR) is better suited for shallow targets like buried utilities because of dielectric permittivity contrasts.
- A single method rarely tells the whole story. Combining complementary methods (e.g., seismic for structure and electrical resistivity for composition) provides independent constraints on subsurface conditions.
- Define survey parameters to match the target:
- Station spacing controls lateral resolution. Tighter spacing resolves smaller features but takes more time.
- Sampling interval (temporal) must satisfy the Nyquist criterion to avoid aliasing: sample at least twice the highest frequency of interest.
- Recording time must be long enough for signals from the deepest target to reach the receivers.

Logistics and Data Acquisition Planning

Even a scientifically sound design fails if it can't be executed in the field.

- Select equipment rated for actual site conditions (e.g., waterproof housings for wetlands, ruggedized instruments for rocky terrain, high-temperature-rated electronics for volcanic areas).
- Design the survey grid to provide optimal coverage while accounting for access routes, topographic obstacles, and potential interference sources.
- Establish clear data acquisition protocols before going to the field. These should cover measurement procedures, instrument settings, quality control checks, and documentation requirements. Consistency across the survey is what makes the data interpretable.

Factors in Survey Design

Geological and Site Considerations

Geology drives method selection. You need to understand (or at least estimate) the subsurface structure, lithology, and physical property contrasts before choosing your approach.

- Characterize the target's expected depth, geometry, and composition, along with the properties of the overburden and bedrock. Key properties include electrical resistivity, seismic velocity, density, and magnetic susceptibility.
- If the physical property contrast between the target and its surroundings is small, you'll need a more sensitive method or denser sampling to detect it.

Site conditions impose practical constraints that directly affect data quality:

- Topography: Steep slopes complicate equipment deployment and require terrain corrections (especially for gravity surveys). Elevation changes also affect seismic travel times.
- Vegetation: Dense forest or brush can limit access and interfere with GPS positioning.
- Cultural noise: Power lines generate electromagnetic interference that degrades magnetotelluric or EM data. Pipelines create magnetic anomalies. Road traffic produces seismic noise. Identify these sources during the planning stage so you can design around them.

Project Objectives and Constraints

Clear objectives are non-negotiable. "Characterize the subsurface" is too vague. Specific objectives look like:

- Map the depth to bedrock across a 500 m × 500 m site
- Detect and delineate a contamination plume at 5–20 m depth
- Identify fault zones within a sedimentary sequence to 200 m depth

The required depth of investigation and spatial resolution follow directly from these objectives. Detecting a 1 m wide void at 3 m depth demands very different parameters than mapping regional basement structure at 2 km depth.

Practical constraints shape every survey:

- Budget limits the number of methods, the survey extent, and crew size.
- Time may restrict how much data you can collect per day.
- Equipment and expertise availability can rule out certain methods entirely.

Optimization means prioritizing: focus the highest-resolution (and most expensive) methods on the most critical areas, and use reconnaissance-level methods elsewhere.

Comprehensive Survey Planning

Defining Objectives and Selecting Methods

A systematic planning process keeps the survey focused and efficient:

1. Define the project objectives and identify the geophysical target(s), including expected depth, size, and relevant physical properties.

2. Evaluate candidate methods against the target characteristics, site conditions, and logistical requirements. Consider each method's strengths, limitations, depth capability, and resolution.

3. Select a primary method and, where the budget allows, one or more complementary methods for independent verification.

4. Estimate the cost, time, and personnel requirements for each method to confirm feasibility.

Survey Geometry and Data Acquisition

Survey geometry determines what you can and cannot resolve.

- Survey area should extend beyond the expected target boundaries. Edge effects and the need for adequate spatial context mean the survey footprint is always larger than the target itself.

- Grid layout: Use 2D lines for simpler targets or reconnaissance. Use 3D grids for complex structures where out-of-plane effects would compromise 2D interpretation.

- Line orientation matters for anisotropic targets. Survey lines should run perpendicular to the expected strike of geological features (e.g., faults, bedding contacts) to maximize the anomaly signature.

- Station spacing should be no more than half the smallest feature you need to resolve (spatial Nyquist criterion). For example, resolving a 10 m wide target requires station spacing of 5 m or less.

Data acquisition parameters to specify:

- Source-receiver configurations (offsets, azimuths)

- Sampling interval and recording time/window

- Number of stacks or repeat measurements for signal-to-noise improvement

- Source parameters (energy, frequency content)

Documentation and Quality Control

Thorough documentation makes the difference between data that can be reprocessed years later and data that's effectively lost.

- Record the full survey plan: methods, parameters, equipment serial numbers, station coordinates, and any deviations from the plan.

- Include maps, diagrams, and parameter tables. Anyone picking up the documentation should be able to reproduce the survey.

- Build quality control into the daily workflow:

- Equipment checks each morning (battery levels, sensor calibration, GPS lock)

- Calibration measurements at reference stations to track instrument drift

- Repeat measurements at selected stations (typically 5–10% of total) to quantify repeatability

- Field data review at the end of each day to catch problems early

- Plan for data backup: copy data to at least two independent storage devices daily. Establish a protocol for secure transfer and long-term archiving.

Limitations and Errors in Design

Method Limitations and Site Effects

Every geophysical method has inherent limitations, and ignoring them leads to uninterpretable results.

- Seismic methods struggle in areas with high attenuation (e.g., thick dry soil, fractured rock) or where velocity inversions cause hidden layers that refraction methods cannot detect.
- Electrical resistivity methods lose sensitivity in extremely resistive environments (e.g., thick dry sand) where current injection is difficult, and in highly conductive environments where current concentrates near the surface.
- GPR penetration depth drops sharply in clay-rich or saline soils due to high conductivity.
- Gravity and magnetics measure potential fields, so solutions are inherently non-unique: different subsurface models can produce the same surface anomaly.

Site effects compound these limitations. Topographic relief requires corrections in gravity, magnetics, and seismic processing. Cultural noise sources (power lines, pipelines, traffic) can overwhelm the signal from the target if not accounted for during planning.

Survey Geometry and Parameter Selection

Errors in survey design are often more damaging than errors in processing, because you can't recover information that was never collected.

- Spatial aliasing occurs when station spacing is too coarse to sample the target's anomaly. The feature either disappears from the data or appears as a false artifact.
- Incorrect assumptions about target depth or properties can lead to choosing the wrong method or wrong parameters entirely.
- Survey orientation affects detection capability. Lines parallel to a fault's strike will show minimal anomaly, while lines perpendicular to strike maximize the signal.
- There's always a trade-off between coverage and resolution. Wider station spacing covers more ground but misses fine detail. Denser spacing resolves small features but covers less area for the same effort.

Mitigation Strategies

Good survey design anticipates problems and builds in safeguards:

1. Redundant measurements: Collect reciprocal measurements (swap source and receiver positions) in resistivity surveys, or repeat measurements at control stations, to quantify data quality and identify errors.
2. Multiple methods: Using two or more independent methods reduces the risk that a single method's blind spot causes you to miss the target.
3. Adaptive design: Review preliminary data in the field. If early results reveal unexpected features or poor data quality, adjust parameters, shift survey lines, or expand coverage before demobilizing.
4. Data processing safeguards: Apply filtering, stacking, and terrain corrections to suppress noise and artifacts. Inversion techniques can help extract subsurface models, but always assess whether the results are geologically reasonable.

5. Pilot surveys: For large or expensive campaigns, run a small pilot survey first to test whether the chosen methods and parameters actually detect the target under real site conditions.

13.2 GPS and positioning techniques

GPS Principles and Applications

GPS and positioning techniques provide the spatial framework for all geophysical fieldwork. Without accurate coordinates and elevations tied to each measurement, even the best gravity, magnetic, or seismic data loses much of its value. These methods range from standard GPS (meter-level accuracy) to carrier-phase techniques like RTK that deliver centimeter-level precision.

GPS Fundamentals

GPS (Global Positioning System) is a satellite-based navigation system providing positioning, navigation, and timing services worldwide. The constellation consists of satellites orbiting at approximately 20,200 km altitude, each transmitting radio signals that encode the satellite's position and the precise time of transmission.

A GPS receiver determines its location through trilateration:

1. The receiver picks up signals from multiple satellites and measures the time delay between transmission and reception.

2. It multiplies each time delay by the speed of light to calculate the pseudorange (distance) to each satellite.

3. With distances from at least four satellites (three for position, one additional to solve for receiver clock error), the receiver computes its 3D position.

Note the distinction: trilateration uses distance measurements, while triangulation uses angle measurements. GPS relies on trilateration.

Differential GPS and Applications

Differential GPS (DGPS) improves on standard GPS by using fixed, ground-based reference stations at known coordinates. Each reference station compares its GPS-derived position to its true position and calculates correction values. These corrections are then broadcast to nearby DGPS receivers, which apply them to reduce systematic errors in the satellite signals.

GPS serves several roles in geophysical surveys:

- Spatial referencing: Every gravity, magnetic, or seismic measurement gets tagged with precise coordinates, enabling the creation of spatially accurate maps and subsurface models.

- Real-time navigation: Surveyors can follow pre-planned survey lines or grids with GPS guidance, ensuring consistent data coverage and minimizing gaps or overlaps.

- Efficiency: GPS-guided surveys are significantly faster than traditional optical surveying methods that require manual angle and distance measurements at each station.

Positioning Techniques for Data Acquisition

Static and Kinematic GPS Positioning

Static GPS positioning places a receiver at a fixed location for an extended observation period (hours to days). The long occupation time allows errors to be averaged out, producing high-accuracy coordinates. This technique is commonly used to establish control points or benchmarks that anchor an entire survey network.

Kinematic GPS positioning tracks a moving receiver in real time, enabling continuous data acquisition along survey lines or profiles. The receiver is typically mounted on a vehicle, boat, or aircraft, or carried by a surveyor on foot. This approach suits surveys requiring dense, continuous measurements over large areas.

Real-Time Kinematic (RTK) and Post-Processed Kinematic (PPK) GPS

RTK GPS achieves centimeter-level accuracy in real time through the following setup:

1. A base station is established at a known location and continuously tracks satellite signals.
2. The base station computes corrections by comparing its known position to its GPS-derived position.
3. These corrections are transmitted via radio link (or cellular network) to the mobile rover receiver.
4. The rover applies the corrections on the fly, resolving carrier-phase ambiguities to achieve centimeter-level positioning.

PPK GPS follows a similar principle but without the real-time radio link. Both the base station and rover record raw satellite observations independently. After the survey, the data are combined and processed in software to resolve carrier-phase ambiguities and compute high-accuracy positions.

RTK vs. PPK trade-offs: RTK gives you positions in the field immediately, which is useful for quality control during the survey. PPK offers more flexibility since it doesn't depend on maintaining a continuous radio link, and post-processing can sometimes recover positions that RTK would have lost due to brief signal interruptions.

Both RTK and PPK use carrier-phase measurements, which track the phase of the GPS carrier signal itself rather than just the modulated code. Carrier-phase measurements are far more precise than code-based pseudoranges, but they introduce integer ambiguities that must be resolved to achieve full accuracy.

Precise Point Positioning (PPP) and Inertial Navigation Systems (INS)

Precise Point Positioning (PPP) achieves high accuracy using a single receiver, with no base station required. Instead, it relies on precise satellite orbit and clock corrections provided by organizations like the International GNSS Service (IGS). PPP typically delivers decimeter-level accuracy with short observations and can approach centimeter-level with longer sessions (30+ minutes). The convergence time needed to reach full accuracy is one of PPP's main drawbacks compared to RTK.

Inertial Navigation Systems (INS) use accelerometers and gyroscopes to track changes in position, velocity, and orientation. On their own, INS solutions drift over time because small measurement errors accumulate. However, when integrated with GPS (GPS/INS integration), the two systems complement each other:

- GPS provides absolute position fixes that correct INS drift.

• INS provides continuous, high-rate positioning that bridges GPS outages (e.g., in tunnels, dense forest canopy, or urban canyons).

This integration is particularly valuable for airborne geophysical surveys where continuous, smooth positioning is critical.

GPS Accuracy and Limitations

Factors Affecting GPS Accuracy

Several factors determine how accurate a GPS position will be:

• Satellite geometry: Satellites spread widely across the sky yield better accuracy. When visible satellites cluster in one part of the sky, the solution degrades. This is quantified by Dilution of Precision (DOP) values; lower DOP means better geometry.

• Atmospheric delays: The ionosphere (charged particles at ~80–1,000 km altitude) and troposphere (lowest ~12 km, where weather occurs) both slow GPS signals, introducing range errors.

• Multipath: GPS signals reflecting off buildings, rock faces, or vehicles before reaching the receiver create interference with the direct signal, distorting range measurements. This is especially problematic in urban areas and near steep terrain.

• Receiver quality: Antenna design, number of tracking channels, and signal processing algorithms all affect measurement precision.

Accuracy Levels of Different GPS Techniques

Technique	Typical Accuracy	Base Station Required?	Real-Time?
Standard GPS	5–10 m	No	Yes
DGPS	1–3 m	Yes (or correction service)	Yes
RTK	1–2 cm horizontal, ~3 cm vertical	Yes	Yes
PPK	1–2 cm horizontal, ~3 cm vertical	Yes (data post-processed)	No
PPP	Decimeter to centimeter	No	Depends on service

RTK and PPK achieve their high accuracy through carrier-phase measurements, which resolve the wavelength of the GPS signal (~19 cm for L1) down to a fraction of a cycle. Standard GPS and DGPS rely on code-based pseudoranges, which are inherently less precise.

Environmental Limitations and Challenges

• Signal blockage: Dense forest canopy, buildings, or steep terrain can obstruct the line of sight to satellites, reducing the number of usable signals or causing complete loss of lock.

• Ionospheric delays: Vary with solar activity, time of day, and geographic location. Dual-frequency receivers (tracking both L1 and L2 signals) can largely eliminate ionospheric errors because the delay is frequency-dependent.

• Tropospheric delays: Caused by variations in temperature, pressure, and humidity. These are modeled using empirical atmospheric models, though residual errors remain, especially for the wet (humidity) component.

- Baseline length: For RTK and PPK, accuracy degrades as the distance between base and rover increases, because the two receivers experience increasingly different atmospheric conditions. Baselines beyond ~20 km typically require network RTK solutions or PPP.

GPS Data Integration for Analysis

Combining GPS and Geophysical Data

GPS coordinates are recorded simultaneously with geophysical measurements, creating spatially referenced datasets. This spatial referencing is what transforms raw measurements into interpretable maps and models of the subsurface.

For example, a gravity survey produces a list of gravity values that are meaningless without knowing exactly where each reading was taken. GPS provides the latitude, longitude, and (critically) the elevation for each station, enabling the computation of gravity anomalies and the construction of contour maps.

Terrain Corrections and Sensor Positioning

GPS elevations play a direct role in several geophysical corrections:

- Gravity surveys: Free-air and Bouguer corrections both require accurate station elevations. Terrain corrections account for the gravitational effect of topography near each station. Even small elevation errors (a few centimeters) can introduce measurable errors in high-precision gravity work, since the free-air gradient is approximately 0.3086 mGal/m.
- Seismic surveys: GPS determines the exact position of each geophone and source point. Accurate positions are essential for proper stacking, migration, and imaging of subsurface reflectors.
- Magnetic surveys: GPS ensures measurements are taken at consistent intervals along survey lines and provides the coordinates needed to produce accurate magnetic anomaly maps and remove regional field trends.

Data Integration and Interpretation

Integrated GPS and geophysical datasets are analyzed using tools such as:

- GIS software for visualizing, querying, and overlaying spatially referenced data layers (e.g., gravity anomalies draped on topography).
- Geophysical modeling software for building 2D or 3D models of subsurface structures, constrained by the spatial coordinates GPS provides.

This integration supports applications ranging from mineral exploration and hydrocarbon prospecting to groundwater studies, geotechnical site characterization, and environmental monitoring. The spatial context GPS provides is what ties geophysical anomalies to real-world locations and geological features, making interpretation and decision-making possible.

13.3 Geophysical instrumentation and data acquisition systems

Geophysical Instrumentation Types and Applications

Geophysical instruments measure physical properties of the Earth and convert them into signals you can record and interpret. Each instrument type targets a different physical property, so choosing the right one depends entirely on what you're trying to detect and how deep it sits.

Magnetometers for Magnetic Surveys

Magnetometers measure the strength and direction of magnetic fields. They're used to map subsurface geology, detect buried objects (pipelines, archaeological artifacts), and study variations in the Earth's magnetic field.

The three main types differ in how they sense the field:

- Fluxgate magnetometers use a ferromagnetic core driven in and out of saturation by an AC signal. The presence of an external field creates an asymmetry in the saturation cycle, producing a measurable output proportional to the field. They measure vector components and are common in borehole and airborne surveys.
- Proton precession magnetometers exploit the precession of hydrogen protons in a fluid (often kerosene or water) around the ambient magnetic field. The precession frequency is directly proportional to total field strength. They give absolute scalar measurements and need no orientation, but they sample relatively slowly.
- Optically pumped (cesium or potassium vapor) magnetometers use laser or lamp excitation to align atomic spin states. They offer the highest sensitivity (down to ~0.001 nT) and fast sampling rates, making them the standard for high-resolution airborne surveys.

Gravimeters for Gravity Surveys

Gravimeters measure variations in the Earth's gravitational acceleration. Because density contrasts in the subsurface produce small gravity anomalies, these instruments help identify geological structures (faults, salt domes, sedimentary basins) and explore for oil, gas, and mineral deposits.

- Spring-based (relative) gravimeters, such as the LaCoste-Romberg, measure the extension of a zero-length spring under gravity. They detect differences in gravity between stations, so you need a base station with a known value. Typical precision is around 0.01 mGal.
- Superconducting gravimeters levitate a niobium sphere in a persistent magnetic field generated by superconducting coils. They achieve extremely high precision (~0.001 μGal) and are used for continuous monitoring of tidal signals, post-glacial rebound, and other long-period gravity changes.
- Absolute gravimeters (e.g., falling-corner-cube instruments) measure g directly by timing a free-falling mass with a laser interferometer. They provide an absolute reference and are used to tie relative survey networks together.

Seismometers for Seismic Surveys

Seismometers measure ground motion caused by seismic waves. They're used to map subsurface geological structures, monitor earthquakes, and investigate the Earth's interior.

- Geophones are small, rugged velocity sensors used in active-source exploration surveys. A coil suspended on a spring inside a magnetic housing generates a voltage proportional to ground velocity. They're deployed in large arrays (sometimes thousands) for reflection and refraction surveys.
- Broadband seismometers use force-feedback electronics to extend the frequency response from ~0.01 Hz to 50 Hz or more. They record everything from teleseismic events to local microseismicity and are the backbone of permanent seismic networks.
- Ocean-bottom seismometers (OBS) are self-contained units deployed on the seafloor. They typically include a three-component seismometer plus a hydrophone and are used for marine crustal studies and

offshore exploration.

Ground-Penetrating Radar (GPR) for Shallow Subsurface Investigations

GPR transmits short pulses of high-frequency electromagnetic waves (typically 10 MHz to 2.5 GHz) into the ground and records reflections from interfaces where the dielectric permittivity changes. It's used for mapping utilities, detecting buried objects, and studying soil and rock properties.

Key trade-offs to understand:

- Higher antenna frequencies (e.g., 1.6 GHz) give better resolution but shallower penetration (centimeters to ~1 m).
- Lower antenna frequencies (e.g., 25 MHz) penetrate deeper (tens of meters in dry rock) but with coarser resolution.
- GPR works best in resistive materials (dry sand, limestone, ice). Conductive materials like clay or saltwater attenuate the signal rapidly, limiting depth of investigation.

Systems can be ground-coupled (antenna on or near the surface, better energy transfer) or air-coupled (antenna raised above the surface, allowing faster survey speeds, common for road and bridge deck surveys).

Electrical Resistivity Meters for Resistivity Surveys

Electrical resistivity instruments inject current into the ground through electrodes and measure the resulting voltage to determine subsurface resistivity. They map geological structures, identify aquifers, and detect contamination plumes.

- DC resistivity uses electrode arrays (Wenner, Schlumberger, dipole-dipole) with different spacings to probe different depths. Multi-electrode systems with automated switching (e.g., SuperSting, ABEM Terrameter) enable 2D and 3D imaging through electrical resistivity tomography (ERT).
- Induced polarization (IP) measures the chargeability of the subsurface, which is the tendency of materials to store electrical charge temporarily. IP is particularly useful for detecting disseminated sulfide mineralization and clay content, since these materials show strong polarization effects that pure resistivity measurements would miss.

Electromagnetic (EM) Instruments for EM Surveys

EM instruments measure the electrical conductivity (or its inverse, resistivity) of the subsurface without requiring direct ground contact. They detect conductive anomalies such as ore bodies, groundwater, and clay zones.

- Frequency-domain EM (FDEM) transmits a continuous sinusoidal signal and measures the secondary field at one or more frequencies. Depth of investigation increases with lower frequency. Portable systems like the Geonics EM-31 and EM-34 are widely used for environmental and engineering surveys.
- Time-domain EM (TDEM) transmits a current pulse and then measures the decay of the secondary field after the transmitter shuts off. The later in time you measure, the deeper the information comes from. TDEM is effective for deeper targets and is less sensitive to near-surface heterogeneity than FDEM.
- Magnetotellurics (MT) is a passive method that uses natural EM fields (generated by solar wind interactions with the ionosphere and by lightning) as the source. By measuring orthogonal electric and

magnetic field components at the surface, you can determine resistivity structure to depths of hundreds of kilometers. MT is used for deep crustal and lithospheric studies.

Principles of Geophysical Data Acquisition

Components of Geophysical Data Acquisition Systems

A data acquisition system converts a physical measurement into digital data you can store and process. The signal chain has five main stages:

1. Sensors convert a physical property (magnetic field, ground velocity, electrical potential) into an electrical signal.

2. Signal conditioning electronics amplify the raw signal and apply filters to improve quality. Common filters include low-pass filters (remove high-frequency noise), high-pass filters (remove low-frequency drift), and notch filters (remove specific interference frequencies, often 50 or 60 Hz powerline noise).

3. Analog-to-digital converters (ADCs) sample the conditioned analog signal at discrete time intervals and convert each sample to a numerical value. The bit depth of the ADC determines its dynamic range: a 24-bit ADC provides roughly 144 dB of dynamic range, which means it can resolve very small signals even in the presence of much larger ones.

4. Data storage (hard drives, solid-state drives, flash memory) holds the digital records for later processing.

5. Control software manages acquisition parameters, monitors data quality in real time, and coordinates the overall system.

Synchronization and Timing in Data Acquisition

When a survey uses multiple sensors spread across an area, all of them need a common time reference. Without it, you can't correctly correlate signals between stations.

• GPS timing is the standard approach. GPS receivers provide time stamps accurate to better than 1 μs, which is sufficient for most geophysical methods.

• Timing accuracy matters most for methods that depend on signal arrival times. In seismic surveys, a timing error of 1 ms can translate to meters of depth error. In TDEM, early-time measurements decay rapidly, so even small timing offsets distort the data.

• Sampling rate (samples per second) must satisfy the Nyquist criterion: sample at least twice the highest frequency you want to record. If your seismic signal contains energy up to 250 Hz, you need a sampling rate of at least 500 Hz (typically you'd use 1000 Hz or higher to provide a comfortable margin).

• Record length should be long enough to capture the full signal of interest, including late arrivals or slow decays, while keeping file sizes manageable.

Instrumentation Selection for Surveys

Matching Instrumentation to Survey Objectives

Choosing the right instrument starts with defining what you need to detect. Three factors drive the decision:

• Target depth determines which methods can reach deep enough. GPR might penetrate 1-10 m in typical soils, while MT can image structures hundreds of kilometers deep.

- Resolution requirements set the frequency or electrode spacing you need. Detecting a 0.5 m diameter pipe requires much finer resolution than mapping a 100 m wide fault zone.
- Expected signal strength relative to noise dictates the required instrument sensitivity and dynamic range. A gravity anomaly from a shallow void might be only 0.05 mGal, so you need a gravimeter with precision well below that.

The instrument's bandwidth should cover the frequency range of the target signal. Using a sensor with too narrow a bandwidth means you'll miss part of the signal; too wide a bandwidth lets in unnecessary noise.

Survey Design and Data Acquisition Parameters

Once you've chosen the method and instrument, the survey design controls what you can resolve:

- Sensor spacing determines spatial resolution and depth of investigation. Closer spacing gives finer detail but takes more time and costs more. A common rule of thumb for seismic reflection surveys is that the receiver spacing should be no more than half the smallest lateral feature size you want to resolve.
- Gain settings should be adjusted so the signal fills a good portion of the ADC's range without clipping. Many modern systems use automatic gain control (AGC) or programmable gain.
- Stacking (repeating measurements and averaging) improves the signal-to-noise ratio by a factor of $\sqrt{N}$, where N is the number of stacks. Doubling the number of stacks gives about a 41% improvement in SNR.
- Combining multiple methods often strengthens interpretation. For example, joint inversion of seismic velocity and electrical resistivity data can reduce ambiguity because different physical properties respond differently to the same geological feature.

Thorough documentation of all survey parameters, instrument settings, and field conditions is essential. Without it, processing and interpretation become unreliable, and the survey may not be reproducible.

Troubleshooting and Maintaining Geophysical Systems

Calibration and Preventive Maintenance

Geophysical instruments drift over time and with changing environmental conditions (temperature, humidity, pressure). Regular calibration keeps measurements traceable and accurate.

- Calibration compares instrument output against a known reference. For magnetometers, this might mean using a Helmholtz coil to generate a known field. For gravimeters, you revisit a base station with a known absolute gravity value. For seismometers, calibration pulses test the frequency response and sensitivity.
- Preventive maintenance includes cleaning connectors, replacing batteries before they degrade, updating firmware, and inspecting cables for damage. These steps reduce the chance of failure during fieldwork, when repairs are difficult.
- Follow manufacturer-recommended maintenance schedules, but also adapt based on field conditions. Equipment used in dusty, humid, or extreme-temperature environments needs more frequent attention.

Troubleshooting Common Issues

When something goes wrong in the field, a systematic approach saves time:

1. Check power first. Weak or failing batteries cause erratic behavior that can mimic sensor problems.

2. Inspect cables and connectors. Intermittent cable breaks are one of the most common field failures, especially at connector junctions where repeated bending occurs.

3. Monitor signal quality indicators. Most acquisition software displays noise levels, signal amplitude, or waveform previews. A sudden change in these indicators points to the failing component.

4. Compare with expected results. If one channel in a multi-channel system shows anomalous data while others look normal, the problem is likely isolated to that sensor or cable.

5. Swap components to isolate the fault. Replace the suspect sensor or cable with a known-good spare and see if the problem follows the component or stays with the channel.

Proper grounding and shielding reduce electromagnetic interference (EMI). In areas near power lines or radio transmitters, EMI can overwhelm weak geophysical signals. Shielded cables, grounded instrument housings, and notch filters at the powerline frequency all help.

Always carry spare parts, backup batteries, and a basic field repair kit. Downtime during a survey is expensive, and remote field sites rarely have electronics shops nearby.

Training and Documentation

Equipment is only as good as the people operating it. Field personnel should be trained on:

- The physical principles behind each method (so they can recognize when data looks wrong)
- Practical instrument handling, including setup, teardown, and transport
- Troubleshooting procedures specific to the equipment being used
- Proper documentation practices

Field logs should record instrument serial numbers, calibration dates, acquisition parameters, environmental conditions (temperature, weather, nearby cultural noise sources), and any anomalies encountered. Digital metadata embedded in data files should be supplemented with written notes. These records are critical for quality control during processing and for anyone who revisits the data months or years later.

13.4 Quality control and data management in geophysical surveys

Quality Control in Geophysical Data

Implementing Quality Control Procedures

Quality control (QC) in geophysical surveys isn't something you tack on at the end. It needs to be embedded in every stage, from data acquisition through processing, to catch problems before they compound.

During data acquisition, QC means actively monitoring instrument performance. This includes:

- Running calibration checks at regular intervals (e.g., beginning and end of each survey line, or at fixed time intervals)
- Monitoring ambient noise levels to flag periods when environmental interference is too high for reliable measurements

- Performing real-time data validation so the field crew can spot dropouts, spikes, or drift while they can still re-collect data

During data processing, QC shifts to verifying that corrections and filters are applied correctly:

- Apply appropriate filters for known noise sources (e.g., 50/60 Hz powerline filtering in EM surveys)
- Correct for known artifacts such as instrument drift, tidal effects (in gravity), or diurnal variations (in magnetics)
- Cross-validate processed results against independent datasets, borehole logs, or other ground truth information

Standardized QC protocols and checklists help keep these steps consistent across different crews and survey days. Regular communication between field crews and processing teams is critical so that acquisition problems get flagged and resolved quickly, rather than discovered weeks later.

Benefits of Robust Quality Control

- Early error detection prevents costly rework. A sensor drift caught on day one saves you from reprocessing (or re-acquiring) an entire survey
- Higher confidence in interpretations because you can demonstrate that known error sources have been addressed systematically
- Better data integration across surveys or projects, since consistent QC documentation makes it possible to compare datasets collected at different times or by different teams
- Reduced risk of misinterpretation, which matters especially when geophysical results inform engineering decisions, resource estimates, or hazard assessments

Sources of Error in Geophysics

Instrumental and Environmental Factors

Geophysical data always contain some degree of error and artifact. Recognizing where these come from is the first step toward managing them.

Instrumental noise includes electronic interference, sensor drift over time, and outright component malfunctions. For example, a magnetometer with a slowly drifting sensor will produce a gradual trend across your survey that has nothing to do with geology. Repeat base station readings help you identify and correct for this.

Environmental factors are often the trickiest to control:

- Weather conditions (wind vibration on geophones, temperature effects on electronics)
- Surface topography creating geometric distortions in seismic or GPR data
- Cultural noise sources like power lines, buried pipelines, fences, or nearby vehicle traffic generating electromagnetic or vibrational interference

Survey design problems can also degrade data quality. Inadequate spatial sampling leads to aliasing, where features in the subsurface appear at incorrect positions or wavelengths. Poor sensor-ground coupling (e.g., a geophone not firmly planted) reduces signal fidelity. Incorrect instrument settings, like choosing the wrong gain or sampling rate, can clip signals or miss important frequencies entirely.

Processing and Interpretation Challenges

Processing itself can introduce artifacts if you're not careful:

- Over-aggressive filtering can remove real geological signal along with the noise
- Incorrect assumptions about subsurface properties (e.g., assuming a constant velocity model in seismic processing when velocities vary laterally) produce distorted images
- Numerical instabilities in inversion algorithms can generate features in your model that don't correspond to anything real

The key habit is to document every potential error source and artifact you've identified, along with what you did (or couldn't do) to mitigate it. This documentation is essential for anyone interpreting the final results. When you encounter complex or ambiguous artifacts, consulting experienced geophysicists can save significant time and prevent misinterpretation.

Data Management for Geophysical Surveys

Organizing and Storing Geophysical Data

A single geophysical survey can generate gigabytes of raw data, processed volumes, and derivative products. Without a clear organizational system, finding and using that data becomes increasingly difficult over time.

File naming and directory structure should follow a standardized convention established before the survey begins. A typical approach encodes the survey name, date, line number, and data type into the filename (e.g., SurveyA_20240315_Line012_rawmag.csv). Consistent naming makes automated processing scripts possible and prevents confusion when multiple people access the same dataset.

Centralized data repositories or database systems provide:

- Secure storage with access controls (so only authorized users can modify raw data)
- Version control to track what changed, when, and by whom
- Efficient search and retrieval across large datasets

Metadata documentation is just as important as the data itself. At minimum, record survey parameters (line spacing, station interval, coordinate system), instrument specifications (model, serial number, firmware version), processing steps applied, and data quality indicators. Without this metadata, even perfectly collected data loses much of its value because future users can't assess its reliability or reproduce the processing.

Data Backup and Sharing Practices

Data loss from a hard drive failure or accidental deletion can be catastrophic if no backup exists. Follow the 3-2-1 rule as a baseline: keep at least 3 copies of your data, on 2 different storage media, with 1 copy stored off-site or in the cloud.

Additional best practices:

- Run data validation checks at each stage of the data lifecycle (acquisition, transfer, processing, archiving) to confirm nothing was corrupted or lost
- Establish clear data sharing policies before the project starts, specifying who can access what data, under what conditions, and in what formats

- Use secure transfer methods (encrypted connections, checksums to verify file integrity) when sharing data with collaborators or stakeholders
- Follow any applicable data privacy or confidentiality requirements, particularly for commercial or government-funded surveys

Data Integrity and Reproducibility in Geophysics

Comprehensive Documentation Practices

Reproducibility means that someone else, given your raw data and documentation, could follow your workflow and arrive at the same results. This is a cornerstone of credible geophysical work.

To achieve this:

1. Document field procedures in detail, including instrument settings, station coordinates, environmental conditions, and any deviations from the planned survey design
2. Archive both raw and processed data in standardized, open formats (e.g., SEG-Y for seismic, GeoTIFF for gridded data) alongside all associated metadata
3. Use version control systems (such as Git) for processing scripts and analysis code, so you can track changes and reproduce any specific version of your workflow
4. Record all processing parameters, including filter settings, inversion parameters, and any manual edits made to the data

Transparency and Adherence to Standards

Every geophysical interpretation carries assumptions, limitations, and uncertainties. Documenting these openly is not a weakness; it's what allows others to use your results appropriately.

- State assumptions explicitly (e.g., "a uniform half-space resistivity of 100 $\Omega \cdot m$ was assumed for the initial model")
- Note known limitations, such as areas of poor data coverage or frequency bands contaminated by noise
- Adhere to community data standards (e.g., those published by SEG, IAGA, or relevant national geological surveys) to ensure your data can be integrated with other studies
- Participate in community efforts to develop and refine QC guidelines and data management best practices, since these standards evolve as instrumentation and methods advance

Proper documentation and archiving don't just protect your own work. They enable future researchers to build on your results, perform meta-analyses across multiple surveys, and ultimately advance the field's collective understanding of subsurface processes.