When Plates Collide

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Abstract

Compressional and contractional tectonics are of interest to various researchers, from rock mechanics and engineering to those studying the hazards, dynamics, and evolution of plate boundaries. We summarize here the terminology regarding deformation associated with compressional and contractional tectonics. We describe the now largely discarded geosyncline theory, which has its roots in contraction. Today, plate-tectonics is the primary theory for explaining the processes shaping the Earth, including earthquakes, volcanoes, and mountain ranges. We emphasize the importance of subduction zones, the most extensive recycling system on the planet, and suture zones, complex boundaries marking the collision zone between two plates. The effects and hazards associated with convergent and collisional plate boundaries are felt far afield and for long distances.

1 Introduction: Notes about terminology

Compressional tectonics is associated with terminology that will be defined here and in other sections. Rock **deformation** is divided into basic components: **translation** (change position), **rotation** (change orientation), **dilation** (change size passively), **dilatation** (change size in response to an active force), and **distortion** (change shape). In basic terms, compressive forces are directed toward each other (→←) and work to squeeze and shorten rock volumes (Figure 1A). A rock responds to stress (σ), including compressional stress, by changing volume or form. Stress has units of force per area (N/m2 or lb/in2 or Pa, pascals) and is characterized by both a magnitude and an orientation on the surface in which it acts (Figure 1). Deformed rocks result from total (finite) deformation over time, from which the forces and mechanisms that created rock textures or structures are interpreted.

Stress can be **normal** (perpendicular to the surface) or **shear** (parallel). Anderson (1905, 1951) linked the orientation of the causative stress tensor relative to the Earth's surface relation to fault types in the upper, shallower levels of the crust (see reviews in Simpson, 1997; Sorkhabi, 2013). The magnitude of stress may not be the same in all directions and thus is defined as maximum σ1> intermediate σ2> minimum σ3.

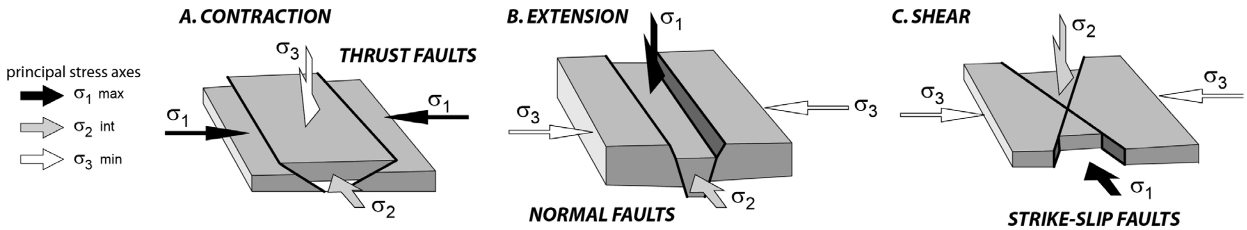

Figure 1. Relationship between stress axes and fault types (after Butler, 2021). **(A)** Rocks displaced by contraction, **(B)** extension, and **(C)** shear. The principal stress axes are identified.

A rock experiences **uniaxial or unconfined compression** when stress is directed toward the center of a rock mass, but more force is applied in one direction, and lateral component forces are zero (σ1 > 0, σ2 = σ3 = 0) (Figure 1A). **Shortening strain** is the change in rock volume due to compressive stress. **Compressional stress** results in shortening features in rocks from the micro to mesoscale, depending on the pressure-temperature (P-T) environment and the nature of the materials comprising the rock.

Rock composition and temperature are critical factors in evaluating how rocks respond to compressional stress. The initial deformation rock experiences during gradually increasing stress is elastic. During this time, changes in stress induce an instantaneous change in sample dimensions as measured by strain. With **elastic** deformation, the strain completely disappears when the stress is removed, and strain is recoverable (Twiss & Moores, 1992). Brittle materials fracture under compressive stress to release stored energy, whereas **ductile** materials deform and compress without failure. Rock layers may fold, or objects change shape, as evidenced by distributed strain. **Plastic** materials flow readily without fracture when the applied stress reaches conditions at or above specific yield stress (Twiss & Moores, 1992).

This book focuses on the processes that occur when the maximum compressive stress is in a horizontal orientation (**contraction**) (Figure 1A). In this case, thrust faulting or folding occurs, shortening and thickening a rock or rock layers. Contraction is also observed as rocks lose volume through crushing, consolidation, or shear. In rock mechanics, contraction is a term that results in a reversible reduction in size, whereas **compression** results in a density increase. Contraction is exposed in the rock record as the shortening of rock layers, thrust or reverse faults, and folds. Thrust faults occur when rocks break along low angles and result in large earthquakes due to the large surface area affected by the process. In this volume, the dynamics of thrust faulting are described by **Pashin et al. (Stratigraphic and Thermal Maturity Evidence for a Break-Back Thrust Sequence in the Southern Appalachian Thrust Belt, Alabama, USA)** and **Cemen and Yezerski (Strain Partitioning in Foreland Basins: An Example from the Ouachita fold-thrust belt Arkoma Basin Transition Zone in Southeastern Oklahoma and Western Arkansas).** Reverse faults result from the rock breaking at high angles in response to compression (Figure 1A). Normal faults occur when the maximum compressive stress is vertical, horizontally extending, and vertically thinning rock (Figure 1B). We cover extensional tectonics in the second volume and strike-slip tectonics (Figure 1C) in the third volume of this series.

2 Setting the Stage: Geosynclinal Theory

The origin of mountains on the Earth has always been debated among philosophers, geographers, and Earth scientists. Since the late 1960s, plate tectonics has been a unifying theory of mountain building (see the next section). Although many theories before plate tectonics were proposed regarding the formation of mountains, one that received wide recognition is the **geosynclinal theory**, commonly attributed to James Hall and his coworkers (Hall, 1859; Dana, 1873; see Fisher, 1978; Frankel, 1982; Friedman, 1999; De Graciansky et al., 2011; Kay, 2014). James Hall based his theory on field observations in the Appalachian Mountains of New York and Pennsylvania, where they observed features characteristic of shallow water sedimentation, such as ripple marks, mud cracks, and shallow-water fossils in sedimentary units that were over 10,000 meters in thickness. But they knew these sediments were deposited in basins where water was only about 100 meters deep. Consequently, Hall proposed that these thick Paleozoic shallow-water sediments must have been deposited in a slowly subsiding basin, receiving a thick succession of shallow-water sediments as it subsided. They coined the term **geosyncline** for this subsiding basin (Figure 2) (Glaessner & Teichert, 1947; De Graciansky et al., 2011). The formation can be further divided into miogeosynclines, eugeosynclines, and orthogeosynclines, depending on the rock strata, location, and nature of the mountain system.

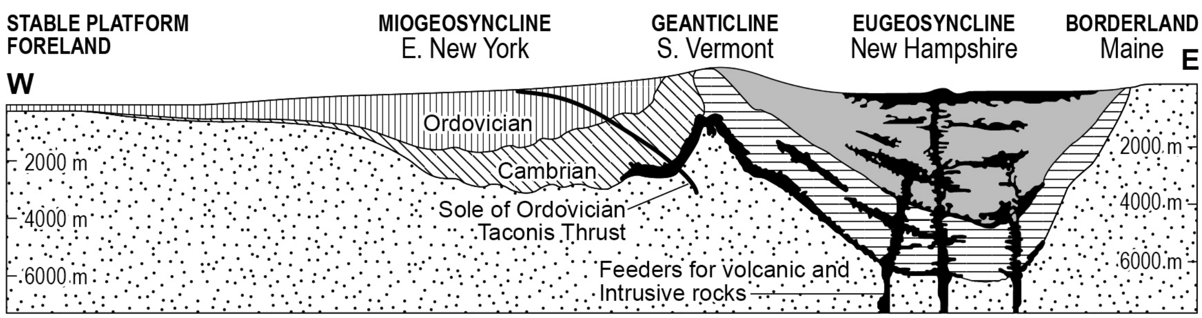

Figure 2. A diagram showing an imagined cross-section of the northern Appalachians prior to the Appalachian Orogeny (after Kay, 1948). A **geanticline** is a ridge that separates two belts of sedimentary rocks. A **eugeosyncline** is a deepwater trough with abundant volcanic rocks and deepwater sediments. A **miogeosyncline** is a basin of mainly shallow water sediments (see De Graciansky et al., 2011).

To explain the deformation that they observed in the Appalachian Mountains, Hall and his coworkers proposed that after thick sediments accumulated, horizontal compressional forces directed from the seaward side of the geosyncline squeezed the sediments, shortened, and thickened the crust, and produced a high-standing mountain chain while pushing much of sediments into the crust. In the 1870s, Dana proposed that the deeply buried sediments melted in high temperature and pressure conditions and generated magma that intruded into the sediments. During the 1890s and early 1990s, geosynclinal theory was widely recognized for explaining the formation of mountain chains, like the Appalachians, Ouachitas, Cordillerans, Urals, Alps, and the Himalayas (Mark, 1992; Şengör, 2021). However, Schaer & Şengör (2008) indicate that the geosyncline theory is not a "made in America" concept. For example, geologists in the Alps had noted the behavior of sediments in deep water basins and ascribed their formation to synclines (e.g., 1828 Elie de Beaumont) (Schaer, 2010).

In 1912, Alfred Wegener published a paradigm-changing hypothesis in his book "The Origin of Continents and Oceans." His hypothesis, called continental drift, suggested that the

Earth's ocean basins and continents changed their positions throughout geological time. Wegener also suggested that all of the continents were together at one time. He called this supercontinent Pangea. Most scientists did not accept Wegener's idea of **continental drift** in the early part of the first half of the 20th century because his lines of evidence were thought to be mostly coincidental. The acceptance of his idea had to wait until the late 1960s, when the data collected from the ocean floor provided evidence that the oceans were indeed temporary: they were opening, closing, and continents were drifting.

Vine & Mathews (1963) worked on magnetic lineations obtained on either side of the mid-Atlantic ridge south of Iceland. They proposed that new oceanic crust is created by the solidification of magma injected and extruded at the crest of a Mid Ocean Ridge (MOR). When this magma cools below the Curie point, ferromagnetic behavior becomes possible, and magnetite in the basalt gets magnetized. The solidified magma (basaltic rocks) acquires a magnetization with the same orientation as the geomagnetic field. They based their hypothesis on the presence of stripes of magnetic anomalies on either side of the MOR. Their findings and those of others who studied the aspects of the geophysical dynamics of MOR gave birth to a unifying theory of Earth Sciences, plate tectonics (see review by Marvin, 2005).

Although geosyncline theory for the evolution of the Earth is today largely discarded, the term is still retained by geologists describing specific basins (e.g., Arabian Gulf geosyncline, Elobaid et al., 2020; Adelaide Geosyncline of South Australia, Preiss, 2000; West Siberian geosyncline, Yolkin et al., 2007). Today, the term is a historical, practical, descriptive, and non-genetic term not meant to be associated with interpretations of a specific tectonic environment (e.g., Preiss, 2000).

3 Plate Tectonics and Compressional Motion

3.1 What are plates?

Plate tectonic theory divides the Earth into rigid layers of crust and upper mantle (**lithosphere**) above the Earth's **asthenosphere**, which can flow at much lower stress levels (Figure 3) (e.g., Anderson, 1995). By their original definition, plates are rigid and include ocean or continental crust or a combination. However, plates do not always correspond with continental margins (e.g., Gordon, 1998). Identifying tectonic plates requires examining geological, geophysical, and geodetic data at multiple sources and scales. These include detailed field mapping and structural analysis, earthquake fault plane solutions, estimates of average rates of plate and fault motion, transform fault azimuths, very long baseline interferometry, satellite laser ranging, Doppler Orbitography and Radiopositioning Integrated by Satellite, and Global Positioning System data (DeMets et al., 2010; Harrison, 2016). Information from these sources helps identify how many plates exist, which has dramatically increased with the technology used to identify them (e.g., n=52, Bird, 2003; n=159, Harrison, 2016). Only 25 **tectonic plates** occupy 97% of Earth's surface (DeMets et al., 2010). The other 3% are **microplates**, defined as relatively small-scale, rigid, geological blocks with a consistent motion or behavior in present-day space with boundaries that behave as plate boundaries (Li et al., 2018). Microplates are located at the major plate boundaries but rotate and behave independently (Hey, 2021). These features may grow into larger plates over time (Seton et al., 2012; Boschman and van Hinsbergen, 2016) or are transient (Hey, 2021).

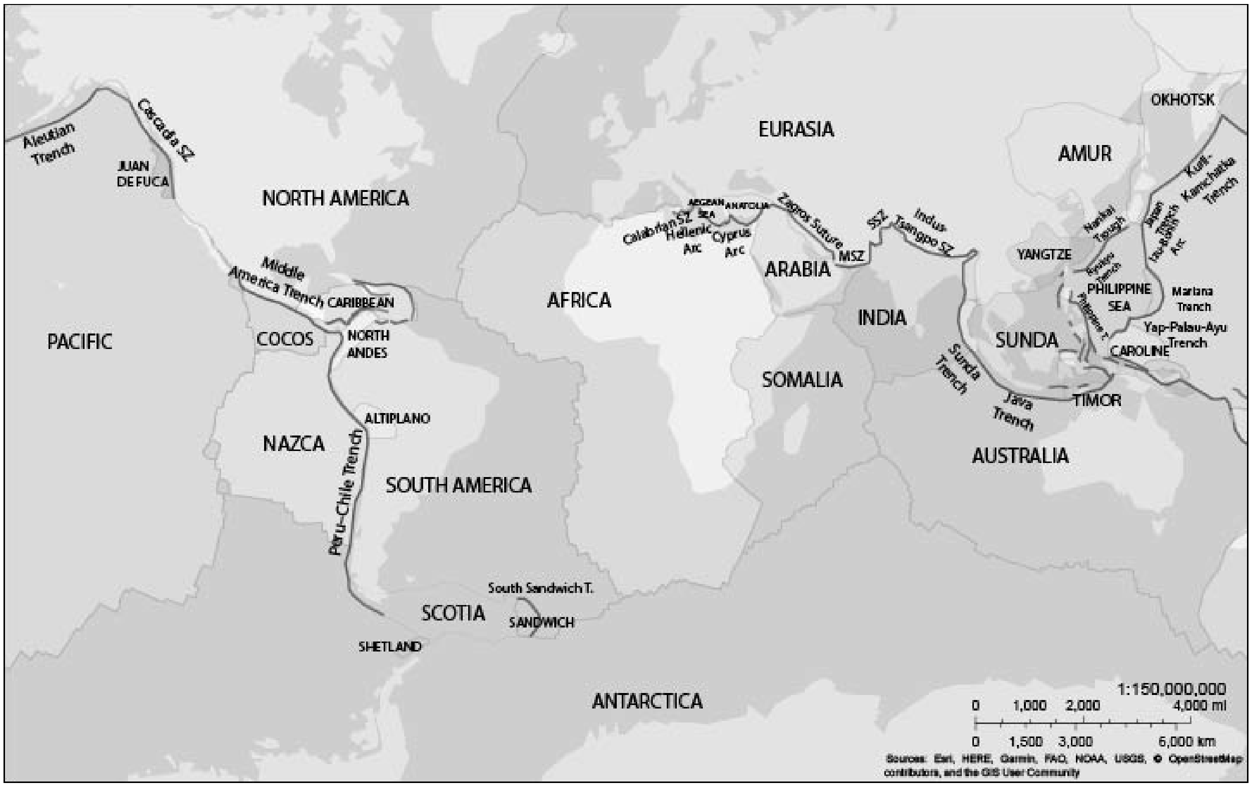

Figure 3. Map of the Earth showing present-day plate configurations and convergent and collisional plate boundaries. Labels are included for some plates and plate boundaries. The map was created using ArcGIS (ESRI) with data from Bird (2003). Convergent and collisional plate boundaries are identified (Coffin et al., 1998). Abbreviations: SZ = suture zone, SSZ = Shyok Suture Zone, MSZ = Makran Suture Zone, Philippine T. = Philippine Trench.

Plates are comprised of **oceanic lithosphere** and/or **continental lithosphere**. The **lithosphere** is the Earth's strong, solid outer shell (Anderson, 1995). The oceanic lithosphere is produced at ocean ridges by decompression melting of upwelling mantle, which cools, thickens, and increases in age as it moves away from ridges (e.g., Condie, 2022). The process creates **mid-ocean ridge basalt (MORB).** This most abundant magma type can be recognized and classified geochemically by source and degree with interaction material recycled in the mantle, spreading rate, and even ocean basin (e.g., Anderson, 1995; Perfit, 2001; Wallace, 2021). The oceanic lithosphere covers ~60% of the Earth's surface (Minshull, 2002; Fowler, 2012), with ocean crust on average 6-8 km thick. Oceanic crust averages 7.1±0.8 km thick away from fracture zones and hot spots and ranges from 5.0-8.5 km (White et al., 1992).

The continental lithosphere is the part of the continental crust and upper mantle that can support long-term geological loads (Anderson, 1995). This layer covers ~40% of the Earth and has a granitic upper portion (32-56 km-thick) underlain by mantle peridotite (96-130 km thick) (DiPietro, 2013). The origin of continental lithosphere differs significantly from mantle lithosphere in that the modification of existing rock creates it through thinning or replacement (Condie, 2005; Sleep, 2005; Eagles, 2020; Şengör et al., 2021). On average, continents are thought mainly to be intermediate (andesitic) in composition with a felsic upper crust and mafic lower crust (Palin et al., 2021). However, based on seismic refraction data, the lower crust may be more felsic in some locations (49-62 wt% SiO_2; Gao et al., 1998; Hacker et al., 2015). This portion of the Earth experiences complex and dynamic interactions that can significantly change its nature, including metamorphism, mixing with mantle-derived melts or other reservoirs, and delamination (e.g., Kay & Mahlburg-Kay, 1991).

Craton lithosphere or **continental platforms** are thick (~200 km) portions of continental thicknesses but differ in age and the mantle dynamics beneath them. Cratons formed during the Archean and platforms are younger features, not underlain by a buoyant mantle that drives convection (Sleep, 2005). Continental lithosphere can thin through extension, orogenic collapse, or underlying mantle processes (e.g., Dewey, 1988; Ruppel, 1995; Lee et al., 2000; Rey et al., 2001; Lavier & Manatschal, 2006;). The subcontinental lithospheric mantle (SCLM) can also be sheared away by cold, shallowly subducting crust, which has an impact on plate buoyancy (e.g., Hernández-Uribe & Palin, 2019) and magmatism (e.g., Wei et al., 2017). Although the oceanic lithosphere assumes the plates are located underwater, some continental lithospheric plates are underwater (e.g., Aegean microplate).

3.2 What are plate boundaries?

Plate boundaries are edges that mark the contact between two plates. Plate boundaries are classified into **divergent** (*extensional*, plates move apart), **conservative** (*strike-slip* if plates slide past each other and *transform* if they also connect divergent plate boundaries), **convergent** (plates move together and a plate is consumed in a subduction zone) or **collisional** (plates move together and plates are joined at a **suture zone**) (see reviews in Cox & Hart, 2009; Le Pichon et al., 2013). Convergent and collisional plate boundaries are classified into a single group (convergent) by most introductory textbooks. These textbooks will also discuss conservative plate boundaries as transform only, with faults classified as strike-slip. Figure 3 highlights the locations of convergent and collisional plate boundaries on Earth as bolder lines, many of which are in the northern hemisphere. Most of Earth's tectonic plates, including many smaller microplates, have a portion in compression (Harrison, 2016).

Although plate boundaries are classified into end-member types, convergent and collisional plate boundaries may also be affected by strike-slip or normal deformation, especially when the plates interact obliquely (Fitch, 1972; Haq & Davis, 1997; Burbidge & Braun, 1998; Bevis & Martel, 2001; Gaidzik & Więsek, 2021). It has long been known that a significant number of plate boundaries have relative velocity vectors that are oblique from normal (>22◦, n=59%) and parallel to the boundary (n=14%) (e.g., Woodcock, 1986). **Composite Transform Convergent (CTC)** plate boundaries define convergent margin plate boundaries that are affected by regional strike-slip faulting along trends that parallel or subparallel the boundary (Ryan & Coleman, 1992). Examples of CTC boundaries may be primarily at **subduction zones** (Figure 4). Subduction zones occur when two lithospheric plates converge, and one plate abruptly descends beneath the other (e.g., Stern & Gerya, 2018; Crameri et al., 2020). CTC boundaries have been identified near **volcanic island arcs** at the Aleutian Ridge and the Philippines (Ryan & Coleman, 1992). Volcanic island arcs are an arcuate continuation of islands with present-day prominent volcanic and seismic activity (Sugimura & Uyeda, 1973). CTCs are present if strike-slip faults develop in the overriding plate (Figure 4) (Beck et al., 1993; McCaffrey, 1993; Bevis & Martel, 2001). The rate of strike-slip faulting in subduction zones is governed by both convergence obliquity and rate (Jarrard, 1986). Normal and strike-slip fault motion in oblique subduction zones have been observed to generate large earthquakes and significantly contribute to its seismic hazards (e.g., Fitch, 1972; McCaffrey, 1996; McCaffrey et al., 2000; Moreno et al., 2008; Melnick et al., 2009; Gaidzik & Więsek, 2021).

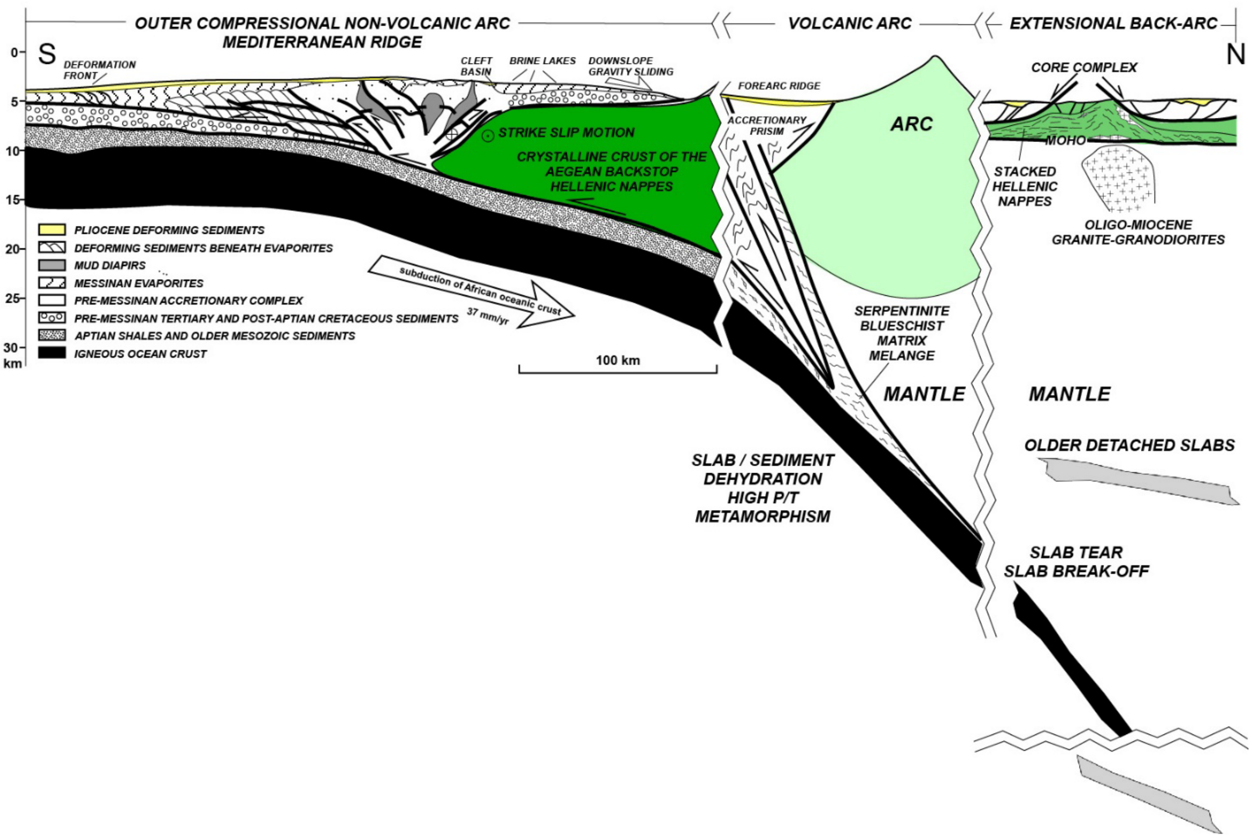

Figure 4. North-south generalized cross-section through the accretionary Hellenic subduction zone showing the structural elements—map of the Mediterranean Ridge after Westbrook & Reston (2002).

Convergent and collisional plate boundaries are characterized by distinct topographical or bathymetric features (Figure 5). Those associated with the oceanic lithosphere will show deep ocean trenches, shallower troughs, ridges of sediment accretion, volcanoes, including seamounts and island arcs, fault lines, and ridges. The US Board on Geographic Names (BGN) Advisory Committee on Undersea Features (ACUF) recommends names of undersea features and official standard names for use in the field or on hydrographic and bathymetric charts. Plate boundaries are often named based on those adopted by the ACUF or by their location, followed by the topographical features they generate (trough, trench, ridge), shape (arc), or nature of deformation (suture, subduction).

However, based on the researcher's focus, the same convergent plate boundary may have several names. For example, the Hellenic subduction zone extends ~1200 km from approximately 37.5°N, 20.0°E offshore of the island of Zakynthos to 36.0°N, 29.0°E offshore of the island of Rhodes (Ganas & Parsons, 2009; Le Pichon et al., 2019). The same feature is sometimes referred to as the Aegean subduction zone (Wortel et al., 1990; Biryol et al., 2011; Crameri et al., 2020), Hellenic arc (Ganas & Parsons, 2009; Royden and Papanikolaou, 2011), or Hellenic arc and trench system (Le Pichon & Angelier, 1979; Papadopoulos et al., 2007). The ACUF assigns the same feature to the Hellenic Trough, Hellenic Trench, or Ionia Basin.

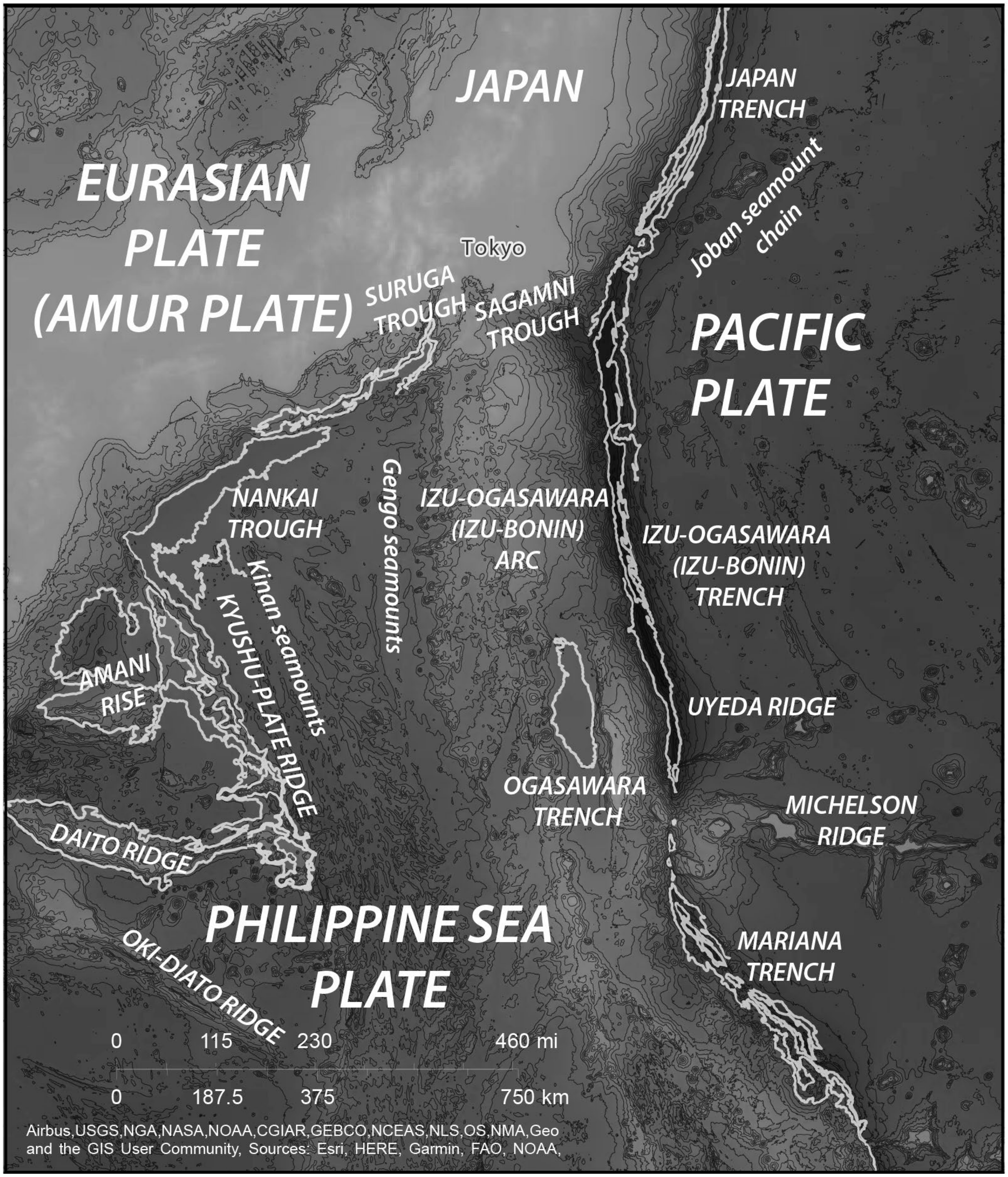

Figure 5. Bathymetry map of subduction zones located near Japan. Some contour lines are highlighted to emphasize particular boundaries and features. The names are after the U.S. Board on Geographic Names (BGN) Advisory Committee on Undersea Features (ACUF).

Trenches, troughs, and **arcs** are often associated with ocean-continent or ocean-ocean subduction zones. Trenches are deeper water regions and exist on the oceanic side of an island arc, whereas a shallow sea exists on the continental side (Figure 4 and Figure 5). Trenches have steep sides like river gorges (e.g., Bellaiche, 1980). Troughs are asymmetrical shallow depressions at the foot of a slope. For example, the Nankai Trough near Japan (Figure 3 and Figure 5) has a maximum water depth that does not exceed 5000 m (Yamano et al., 1984). In contrast, the Izu-Bonin Trench reaches 9780 m (e.g., Bellaiche, 1980). Arcs are curved subduction zones, with the curvature associated with the negative buoyancy and steep dip of the down-going slab (Turcotte & Schubert, 2002), rates of the plate motion, or specific mechanical conditions that govern their geometry (Mahadevan et al., 2010).

3.3 Subduction and suture zones

Subduction zones are considered the most extensive recycling system on the planet and play a key role in Earth's geodynamics and crustal evolution (e.g., Li et al., 2013). The majority of the driving force of plate motion today is generally thought to be slab pull caused by the densification of subducted ocean crust (Forsyth & Uyeda, 1975; Chen et al., 2020; Palin & Santosh, 2021). Subduction zones also form large-scale metal ore deposits (e.g., Sawkins, 1972, Glasby, 1996, Rosenbaum et al., 2005; Kerrich et al., 2005; Li et al., 2013). Igneous activity within these zones forms most of the world's ore deposits (Stern, 2002). These include porphyry copper ± molybdenum ± gold deposits (PCDs), considered the most representative and valuable magmatic-hydrothermal metallogenic systems (Sillitoe, 2010; Rosenbaum et al., 2005; Chen & Wu, 2020). PCDs are located in magmatic-hydrothermal systems in the crust above subduction zones (Sillitoe, 2010; Chen & Wu, 2020; Xue et al., 2021). Here, ore-forming elements are enriched in the mantle wedge due to metasomatism driven by subducting slab-derived fluids (e.g., Zheng, 2019).

Subduction zones are classified based on the fate of ocean basin sediment and detritus accumulated through the erosion of continental and volcanoes that accumulate in the trench or trough (von Huene & Scholl, 1991). A thorough discussion of subduction zone dynamics is provided in this volume by **Agard and coauthors (*Subduction and obduction processes: the fate of oceanic lithosphere revealed by blueschists, eclogites, and ophiolites*). Erosive** subduction zones have crustal sedimentary material removed through subduction, whereas **accretionary** subduction zones show upper plate growth due to frontal accretion or underplating (e.g., von Huene & Scholl, 1991; Clift & Vannuchhi, 2004; Straub et al., 2020). Subduction erosion can still occur beneath accretionary margins and contribute to the geochemistry of arc volcanoes (Clift & Vannuchhi, 2004; Straub et al., 2020).

Convergent plate boundaries are often evident on bathymetry maps based on the subduction of one plate as it is consumed (Figure 5). However, Dewey (1977) noted that **suture zones** that delineate the zones of collision between two continents are rarely simple and rarely create easily recognizable lines (Figure 6). These zones are locations where oceans and back-arc basins are closed (Burke et al., 1977). Their complexity is attributed to the irregular margins of colliding continental plates that generate broad and complex deformation zones (e.g., Chetty, 2017). These locations can involve multiple fault structures, with many experiencing high-strain, intense, and sometimes multi-stage deformation (Abdelsalam & Stern, 1996). P paleolocation of crusts on either side of the zone helps identify such zones, often facilitated by paleomagnetism studies.

As seen in Figure 6, suture zones incorporate a wide range of rock materials. They are critical locations for developing orogenic gold deposits where hydrothermal fluids are localized near and along convergent margins and in the middle and upper crust (e.g., Goldfarb et al., 2001;

Pour et al., 2016). Goldfarb et al. (2001) document numerous goldfields worldwide associated with suture zones over Earth's history. Collision granitoids within suture zones can concentrate economically critical minerals, such as tungsten (scheelite) and gold, rare-metal granites and pegmatite, and colored gemstones (e.g., Koroteev et al., 2009). Although these mountain-building events occur with lower thermal gradients than subduction zone settings and thus are not favorable for the hydrothermal mobilization of ore-forming elements, they are sometimes preceded by subduction zone convergence which provides ample preliminary enrichment before collision (Zheng et al., 2019).

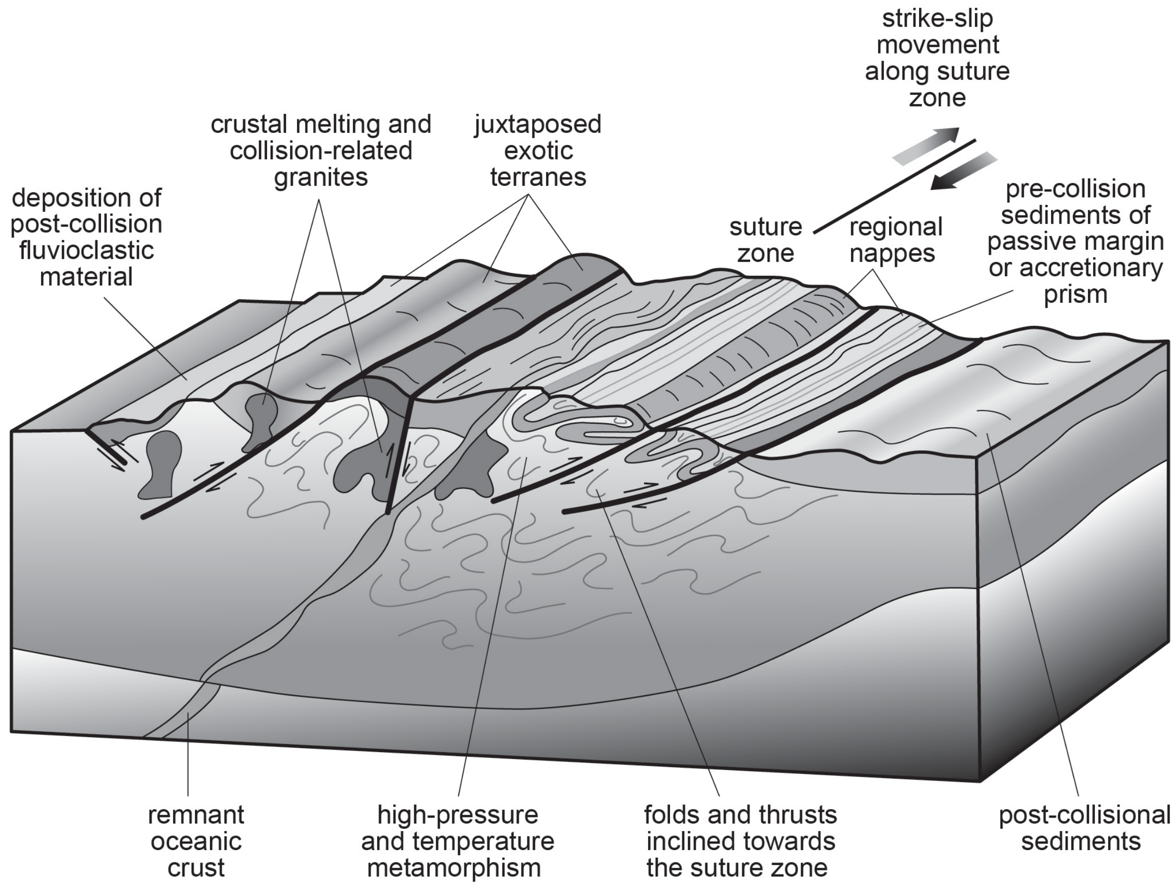

Figure 6. A schematic example of a suture zone. The picture is from the Open University (Geological processes in the British Isles).

Sedimentary rocks in suture zones have recorded multiple facies types attributed to the deep-water ocean's nature to erosion from the overriding continental plate. Shales, turbidites, and deep-water radiolarian chert are recorded in suture zones (e.g., Chakrabarti, 2016). Suture zones can contain chemically and mineralogically matured multicycle sediments (Chetty, 2017). Thick units of sedimentary rocks can be partially subducted under the overriding lithosphere, creating metamorphic assemblages that record the collisional process. Depending on protolith and collision conditions, these metamorphic assemblages can be high-pressure eclogites and Barrovian-grade metapelites. Suture zones are often characterized by high-pressure blueschist–eclogite belts to even ultrahigh-pressure metamorphic (UHPM) complexes, remnants of the subduction zone that existed between two continents (Chetty, 2017).

Various igneous rocks may be present within suture zones, including mafic (ophiolites, serpentinized gabbro, sheared volcanic, blueschists) and felsic assemblages (syn-tectonic high Si, peraluminous granites). Deformed alkaline rocks and carbonatites (DARCS) delineate the boundaries of major Proterozoic suture zones (e.g., Burke et al., 2003; Leelanandam et al., 2006; Catlos et al., 2008). Perhaps the most recognizable feature of suture zones is stratigraphically intact **ophiolites**, remnants of the crust and upper mantle portions of ocean lithosphere or back-arc basins that disappeared between the two continents (e.g., Steinmann, 1906; Hess, 1955; Hawkins, 2003). **Supra-subduction zone (SSZ) ophiolites** are obducted oceanic crust with island arc geochemical characteristics that formed via seafloor spreading (synmagmatic extension) directly above the subducted oceanic lithosphere (Miyashiro, 1973; Pearce et al., 1984; Shervais & Kimbrough, 1985; Hawkins, 2003; Pearce, 2003). Ophiolites in suture zones provide a critical record of deep oceanic crust and ancient seafloor processes (Chetty, 2017).

The timing of collision and convergence of particular subduction and suture zones can be challenging and is often disputed. See a discussion about this topic as it relates to the development in the Himalayas by **Robinson and Martin (*Genesis of Himalayan stratigraphy and the tectonic development of the thrust belt*) and Catlos [(*Records of Himalayan Metamorphism and Contractional Tectonics in the central Himalayas (Darondi Khola, Nepal*)].** For example, although the Himalayan collision is often cited as during the Paleocene (Patriat & Achache, 1984; Klootwijk et al., 1992; Rowley, 1996; Yin & Harrison, 2000; Najman et al., 2001; Ding et al., 2005), much younger constraints are also suggested (e.g., Eocene/Oligocene boundary, Aitchison et al., 2007). Collision may have been a two-stage process, with events occurring in the Paleocene (soft) and Miocene (hard) collision (van Hinsbergen et al., 2012; see review in Parsons et al., 2020). Each component in the suture zone environment has the potential to provide evidence for its history, including the onset of sediment deposition, timing of metamorphism and recrystallization, and paleomagnetic evidence for the locations of the continental block before the collision. Suture zones are often at sites of high topography, but the development of large mountain belts associated with plate convergence occurs significantly after initial contact. In this volume, **Giri and Hubbard (*Lateral heterogeneity in convergent mountain belt settings*)** discuss how orogenic belts worldwide record deformation along strike.

Subduction Zone Initiation (SZI) is the onset of downward plate motion forming a new slab, which later evolves into a self-sustaining subduction zone (Crameri et al., 2020). In this volume, SZI is discussed as relevant to the Eurasian margin by **Bo et al. (*When and why the Neo-Tethys ocean begins to subduct along Eurasian margin: a case study from Iran*)** and along the Hellenic arc by **Catlos and Çemen (*A Review of the Dynamics of Subduction Zone Initiation in the Aegean Region*).** The Hellenic arc (Figure 4) has perhaps the most significant discrepancy between the onset subduction of the African (Nubian) slab beneath the Aegean microplate. Some studies suggest a Cenozoic SZI age, although estimates from the Eocene-Pliocene (e.g., Meulenkamp et al., 1988; Spakman et al., 1988; Papadopoulos, 1997; Brun & Sokoutis, 2010; Le Pichon et al., 2019) to Mesozoic (Late Cretaceous-Jurassic) (Faccenna et al., 2003; van Hinsbergen et al., 2005; Royden & Papanikolaou, 2011; Jolivet et al., 2013; Crameri et al., 2020, van Hinsbergen et al., 2021). Tools used to time SZI are similar to those at suture zones. They include sediment deposition in the accretionary prism (Figure 4), paleomagnetism, the analysis of topography combined with estimates of slab age and depth, reconstructions of subducted slabs using tomography, and the timing of metamorphism and volcanic activity that parallels the subduction zone (e.g., Crameri et al., 2020).

3.4 Hazards associated with compressional plate boundaries

The theory of plate tectonics suggests that plate interaction occurs primarily at the plate boundaries (see review by Gordon, 1998). Plate boundaries are often shown as thin lines and narrow zones (e.g., Figure 3 and Figure 5). However, the effects of convergent and collisional plate boundaries are felt far afield. Figure 7 shows the compressional fault systems associated with convergent and collisional plate boundaries in parts of Europe, the Middle East, and Asia. The effects of these plate boundaries extend far beyond their contact zones. The figure also outlines several **orogenic belts**, which are deformation zones due to horizontal compression, gravity, heat, and climate-driven erosion (DiPietro, 2018). Orogenic belts are explicitly discussed in this volume by **Yilmaz et al. (*Tectonics of Southeast Anatolian Orogenic Belt*).** Orogens not only imply collisional dynamics and the nature of the kinematics in that region, but the term is also a culturally-relative statement that the velocity field in that region has more degrees of freedom than present data constrain (Bird, 2003). Orogenic belts form due to a collage of processes, including magmatism, metamorphism, sedimentation, and deformation (Chetty, 2017). The end stages of orogenic belts are described in this volume by **Foster et al. (*Extensional Collapse of Orogens: A review with an example from the Southern Appalachian Orogen*).**

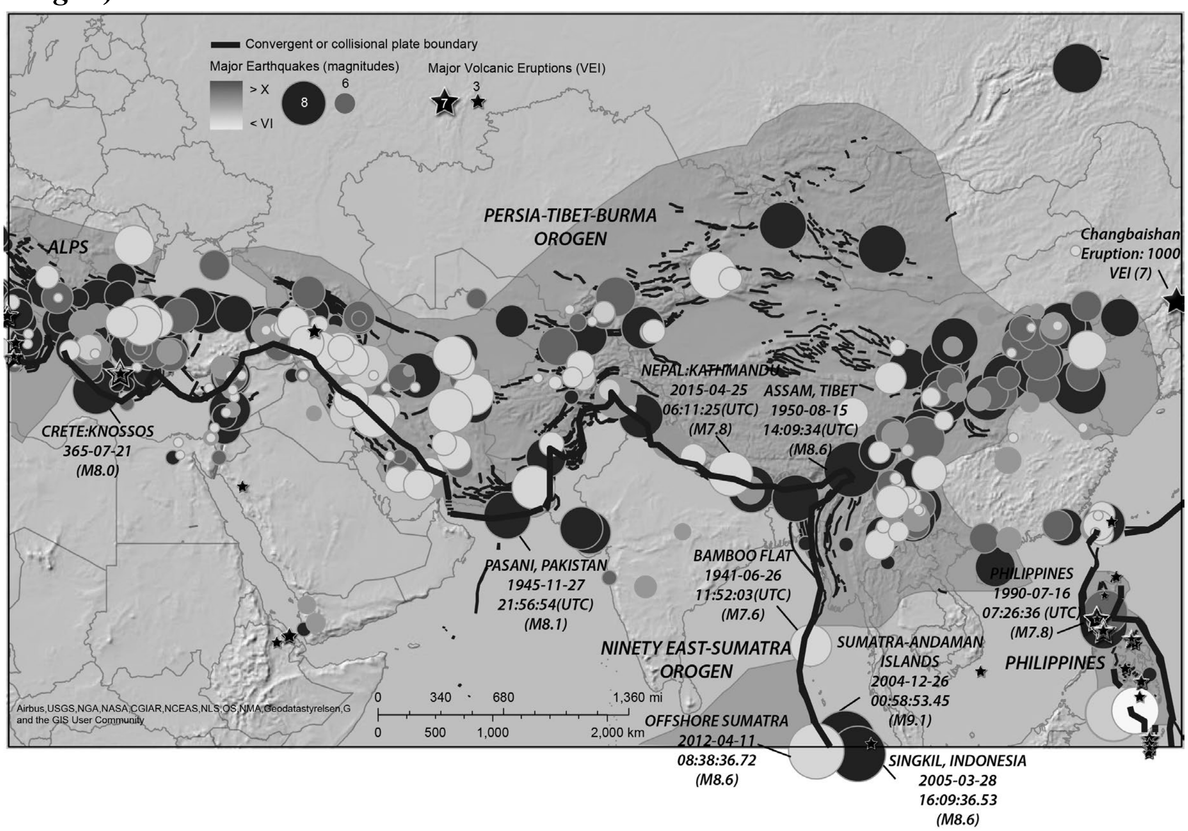

Figure 7. Map (ArcGIS) showing the major collisional and convergent plate boundaries with significant earthquakes and volcanic eruptions overlain. Also included are the boundaries of orogenic belts (Bird, 2002) and fault systems with an element of compression only. Convergent and collisional plate boundaries are identified by Coffin et al. (1998). Global active fault lines from information collected by the Global Earthquake Model Foundation.

Figure 7 shows the relationship between some of Earth's largest earthquakes and destructive volcanoes and convergent and collisional plate boundaries. According to the USGS, all of the Earth's most destructive and largest magnitude earthquakes occurred at convergent or collisional plate boundaries (Table 1). According to Table 1, subduction zones around the Pacific plate account for most of these events, including the Aleutian arc, Japan Trench, Peru-Chile, Columbia-Ecuador, and Kurile-Kamchatka subduction zones. Subduction zones host Earth's most destructive megathrust earthquakes, which are also associated with devastating tsunamis (e.g., Plafker, 1969; Cisternas et al., 2005; McCaffrey, 2008; Melnick et al., 2009; Toda & Tsutsumi, 2013; Bletery et al., 2016). **Tsunamis** are catastrophic wave motions generated by shock waves that cover large parts of the sea and behave intricately in coastal zones (Sugawara et al., 2008). All events in Table 1, except for the 1950 Assam-Tibet earthquake, are **tsunamigenic earthquakes**. Tsunamis triggered by earthquakes are partially generated due to a shallow focus coupled with large rupture areas associated with lower-angle megathrust faulting at subduction zones (e.g., Sugawara et al., 2008; Bilek & Lay, 2018). The largest earthquakes in Table 1 were associated with significant rupture areas: the 1960 Great Chilean Earthquake (Valdivia) at the Peru-Chile trench had a rupture length of 920±100 km (e.g., Cifuentes, 1989), whereas the 1964 Aleutian-Alaska megathrust fault ruptured a length of 600-800 km (Ichinose et al., 2007). The 2004 Sumatra - Andaman Islands earthquake resulted in a rupture length of 1500 km (e.g., Gahalaut et al., 2006).

Table 1. List of Earth's twenty largest earthquakes (source: USGS, 2019) [a]

Location[a]	Day and Time	Lat.	Long.	Mag	Depth	Location
Great Chilean Earthquake (Valdivia)	1960-05-22 19:11:20.00	-38.143	-73.407	9.5	25	Peru-Chile Trench
Prince William Sound (Great Alaska)	1964-03-28 03:36:16.00	60.908	-147.339	9.2	25	Aleutian Subduction Zone
Sumatra - Andaman Islands	2004-12-26 00:58:53.45	3.295	95.982	9.1	30	Sumatra–Andaman Subduction Zone
Great Tohoku Japan	2011-03-11 05:46:24.12	38.297	142.373	9.1	29	Japan Trench
Kamchatka, Russia	1952-11-04 16:58:30.00	52.623	159.779	9	21.6	Kuril-Kamchatka Subduction Zone
Ecuador-Colombia	1906-01-31 15:36:10.00	0.955	-79.369	8.8	20	Colombia–Ecuador Subduction Zone
Quirihue, Chile	2010-02-27 06:34:11.53	-36.122	-72.898	8.8	22.9	Peru-Chile Trench
Rat Islands, Aleutian Islands, Alaska	1965-02-04 05:01:22.00	51.251	178.715	8.7	30.3	Aleutian Subduction Zone
Unimak Island, Aleutian Islands Alaska	1946-04-01 12:29:01.00	53.492	-162.832	8.6	15	Aleutian Subduction Zone

Assam-Tibet	1950-08-15 14:09:34.00	28.363	96.445	8.6	15	Indo-Asia Collision (Mishmi Thrust)
Offshore Sumatra	2012-04-11 08:38:36.72	2.327	93.063	8.6	20	Sumatra–Andaman Subduction Zone
Singkil, Indonesia	2005-03-28 16:09:36.53	2.085	97.108	8.6	30	Sumatra–Andaman Subduction Zone
Adak, Alaska	1957-03-09 14:22:33.00	51.499	-175.626	8.6	25	Aleutian Subduction Zone
Vallenar, Chile	1922-11-11 04:32:51.00	-28.293	-69.852	8.5	70	Peru-Chile Trench
Tual, Indonesia	1938-02-01 19:04:22.00	-5.045	131.614	8.5	25	Banda Sea Arc
Kuril'sk, Russia	1963-10-13 05:17:59.00	44.872	149.483	8.5	35	Kurile-Kamchatka Subduction Zone
Mil'kovo, Russia	1923-02-03 16:01:50.00	54.486	160.472	8.4	15	Kurile-Kamchatka Subduction Zone
Atico, Peru	2001-06-23 20:33:14.13	-16.265	-73.641	8.4	33	Peru-Chile Trench
Sanriku-oki, Japan	1933-03-02 17:31:00.00	39.209	144.59	8.4	15	Japan Trench
Bengkulu, Indonesia	2007-09-12 11:10:26.83	-4.438	101.367	8.4	34	Sumatra–Andaman Subduction Zone

a. Magnitude estimated using the moment magnitude scale (Mw) or Moment W-phase. Some locations are seen in Figure 7.

The 1950 Assam-Tibet earthquake (Figure 7, Table 1) influenced rivers in India, Burma, East Pakistan, Tibet, and China. Many flooded and changed their courses permanently (Ben-Menahem et al., 1974; Mrinalinee Devi & Bora, 2016). Sharma & Zaman (2019) describe the ecological impact of the Assam-Tibet earthquake on the Brahmaputra River as it was affected by liquefaction and contamination by sulfur emanating from underground coal beds and oil seepages. In addition, **seismic seiches** related to the earthquake were recorded in several fjords and lakes over 7000 km away in Norway (Kvale, 1955; McGarr, 2011). Seismic seiches are standing waves in closed or partially closed bodies of water due to the passage of seismic waves from an earthquake (Garr, 2019). Based on a historical assessment, earthquakes in the Himalayan region may not be expected to be as large as those in subduction zones (Srivastava et al., 2013). However, the variations in seismicity of collisional mountain belts are related to a complex interplay between rheology, fault style, kinematics, and the tectonic stress regime, but the parameters that control earthquake behavior in orogenic mountain belts remain unclear (e.g., Dal Zilio et al., 2018).

Ground shaking due to earthquakes at convergent and collisional boundaries often triggers significant mass wasting events, including landslides, rockfalls, and liquefaction. Evidence for giant terrestrial landslides is present along several convergent and collisional plate boundaries worldwide (Mather et al., 2014; Roberts et al., 2014). Landslides develop over steepened slopes and are triggered by large earthquakes or volcanic eruptions. If these events are located near coastal areas, tsunamis can develop. Significant triggers for tsunamis are subaqueous earthquakes and slides (Sugawara et al., 2008). Submarine landslides generated by earthquakes have triggered devastating tsunamis in the Aegean region (e.g., Dominey-Howes, 2002; Okal et al., 2009; Ebeling et al., 2012). The sloping bottom of the Hellenic arc, coupled with thick accumulations and high rates of recent sedimentation, closely spaced active faults, active earthquakes, and **magmatic diapirism** (where less dense rock rises through buoyant forces, Rajput & Thakur, 2016), contribute to its high hazards of tsunamis in the region (e.g., Ferentinos, 1990; Hooft et al., 2017). The eruption of Santorini in 1610 BCE generated a tsunami that affected civilizations throughout the eastern Mediterranean (Dominey-Howes, 2004, Friedrich, 2006, Marinatos, 1939; Hooft et al., 2017). Detailed bathymetry across the Mediterranean is critical in understanding tsunami propagation and mitigating its impacts (e.g., CIESM, 2011).

Figure 7 shows the relationship between convergent plate boundaries and significant volcanic eruptions. The Earth's most extensive volcanic fields in terms of basaltic and silicic eruptions are not found at convergent plate boundaries but are over **large igneous provinces** (LIPS) (e.g., (Coffin & Eldholm, 1994; Bryan & Ernst, 2008; Bryan et al., 2010). However, the origin of LIPS may lie in the subduction process that perturbs mantle dynamics, forces extension in the back-arc region, thins the lithosphere, and trigger large-scale and voluminous basalt eruption (Zhu et al., 2019). The return flow of **slab avalanches** from the mantle transition zone can also generate LIPS (Gurnis, 1988, Coltice et al., 2007; Condie et al., 2021). Slab avalanches develop when large-volume subducted slabs temporarily stagnate within the transition zone and periodically penetrate the lower mantle (e.g., Solheim & Peltier, 1994; Deschamps & Tackley, 2009; Yang et al., 2018). Slab avalanches are controlled by mantle thermal instabilities and accelerate as slab sinking rates increase with time (e.g., Solheim & Peltier, 1994; Yang et al., 2018).

Subduction zones also produce eruptions that are most commonly observed and most dangerous to human populations (Siebert et al., 2015). Subduction zone volcanism propels volcanic gases (e.g., SO_2, CO_2, H_2S) and ash into the stratosphere or troposphere and has affected short-term climate (Bryan et al., 2010; Cooper et al., 2018) and the carbon cycle (Zhu et al., 2021). Some sulfur gases convert to sulfate aerosols in the stratosphere and scatter radiation (e.g., Robock, 2000). The dust veil index (DVI/Emax) measures an eruption's release of dust and aerosols over the years following the event, especially the impact on the Earth's energy balance (Lamb, 1985). For example, the AD 1835 eruption of Volcan Cosiguina, Nicaragua, which is located on a convergent margin where the oceanic crust of the Cocos plate subducts beneath the western edge of the Caribbean plate, is recorded as a volcano has a DVI/Emax of 4000, with ashfall recorded as far as 1900 km away (Scott et al., 2006).

Climate change is intrinsically related to collisional plate boundaries, as topographic barriers interact with the Earth's atmosphere (e.g., Burbank, 1992; Cronin, 2009; Ruddiman, 2013; Song et al., 2021) and subducting slabs at collisional boundaries eliminate megatons of carbon (e.g., Clift, 2017; Plank & Manning, 2019). Controls on the subduction process may be related to climate change (Lamb & Davis, 2003; Iaffaldano et al., 2006). The onset of the Himalayan monsoon is related to India-Asia convergence and is widely studied for understanding the timing of mountain building (e.g., Clift et al., 2008; Allen & Armstrong, 2012; Webb et al., 2017). Mountain ranges are barriers to atmospheric circulation, and exposures of rocks in the mountainous regions can also drive the drawdown of atmospheric gasses through weathering processes that may be directly related to climate change (e.g., Stern & Miller, 2018).

4 Objectives and Organization of the Book

This volume was written to create an up-to-date and relevant compendium valuable reference for Earth Sciences students, including advanced undergraduate and graduate students, postdocs, educators, research professionals, and policymakers in academia and industry. These papers aimed to synthesize current knowledge of complex geological topics surrounding global collisional and convergent plate boundaries with an accessible approach and transparent organization. The papers are meant to be readable for a range of consumers. Several reviewers helped to identify topical oversights and assure that citations fairly represent the body of existing information. The topics are mentioned in the preface, in the text of this introduction, and highlighted in the volume's table of contents.

Acknowledgments, Samples, and Data

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