Subduction

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Synonyms
Convergent plate boundary; Destructive plate boundary; Plate consumption

Definition
At a convergent plate boundary, one of the plates necessarily passes beneath the other. This process, first termed subduction by A. Amstutz in 1951 when discussing the evolution of alpine structures, allows nearly all oceanic lithosphere to be recycled into the mantle. By contrast, only the lower portion of the continental lithosphere (the lithospheric mantle) may, under certain conditions, be transported into the deep mantle. Since the continental crust resists subduction at depths beyond about 100 km due to buoyancy forces, this type of convergence is typically called collision and forms an “orogenic” or a “mountain” belt of which the Alps or the Himalayas are classical examples. Subduction zones can generally be traced at the surface of the Earth because they are associated with large earthquakes, active volcanoes, and sometimes mountain building.

Classification
There are two main types of subduction (Fig. 1), depending on the nature of the downgoing plate (hereafter called slab):

- Oceanic subduction (often just called subduction) when an oceanic plate subducts
- Continental subduction (often just called collision) when a continental plate subducts

Oceanic subduction can last tens or hundreds of millions of years without a significant change in the topography and tectonics in the upper plate, whereas continental subduction, which typically occurs after a period of oceanic subduction, rapidly causes either a plate reorganization or mountain building. Below, we will focus primarily on oceanic subduction processes (see “Orogeny” entry for details on continental subduction).

Each of these subduction types can be divided in two “subtypes” depending on the nature of the overriding plate. The most common situation at a convergent plate boundary is the subduction of an oceanic plate beneath a continental plate (~67 %, 45,000 km, see entry “Active Continental Margin”). Among the remaining 33 %, the converging continents contribute for about half, i.e., 17 % (11,200 km), of the total, intra-oceanic subduction zones for about 15 % (10,000 km) and continents subducting beneath oceanic plates for only 1 % (600 km) (Lallemand et al., 2005a).

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Many other criteria or combinations of criteria have been proposed to describe the wide variety of subduction zones (Jarrard, 1986). The most commonly used, first introduced by Uyeda in 1984, is the opposition between the Chilean-type and the Mariana-type subductions. The Chilean-type is characterized by a young subducting plate with a shallow slab dip, a high degree of interplate coupling, inducing high stresses and producing large earthquakes, mountains, and thus terrigenous sediment supply feeding an accretionary wedge. On the contrary, the Mariana type is characterized by an old subducting plate with a steep slab dip; a low degree of interplate coupling, producing low stresses; and only moderate seismicity, the presence of back-arc extension, and a low topography unable to feed the trench. Despite the correct description of these two end-member subduction zones, Heuret and Lallemand (2005), based on revised global datasets, have shown that some of these associations of parameters are not statistically verified. Those which are statistically verified in modern subduction zones are the combination of upper plate compressional stress (typically Chile), shallow-dipping slabs (less than 50°), and upper plate seaward absolute motion, or, reversely, upper plate extensional stress (typically Mariana), steeply dipping slabs (more than 50°), and upper plate landward absolute motion (Lallemand et al., 2005b). The other characteristics such as plate age, seismicity, or the presence of an accretionary wedge do not show simple correlations.

**Distribution on Earth**

Ninety percent of the oceanic subduction zones are concentrated around the Pacific Ocean and in Southeast Asia (48,700 km). The rest are distributed in the Atlantic Ocean (5,000 km) and the Mediterranean Sea (1,350 km). This asymmetric distribution at Earth’s surface results from the breakup of the Pangea in Late Paleozoic and Early Mesozoic time accompanied by the rifting of the Atlantic and Indian oceans and the closure of the Tethys and Panthalassa oceans. Because the most rapid subduction rates are presently observed in the Western Pacific and Southeast Asia, Xavier Le Pichon in 1990 (lessons at Collège de France) compared this relatively small region with a huge pit.
centered on the Philippine Sea where oceanic plates are recycled into the mantle. Subduction rates generally range between 1 cm/year off Sicily, for example, and 10 cm/year off Japan, except in Southwest Pacific where it exceptionally reaches 24 cm/year offshore the northernmost Tonga archipelago.

**Distinctive Characteristics of Subduction Zones**

The plane forming the interface between converging plates has distinctive properties compared with a “classical” active fault. It accommodates the underthrusting of one plate beneath another. Its dip angle may vary between 10° and 35° for the shallow part which is seismogenic (Heuret et al., 2011), and it can reach 60° in its deepest part. The subduction interface outlines the contact surface between the plates, i.e., from the trench down to the limit with the convecting mantle. The main differences with a “classical” active fault are:

- A great size in both downdip width (typically between one and two hundreds of km) and lateral extent (length between a few hundreds and a few thousands of km).
- A lower thermal gradient along the interface because of the continuous advection of low-temperature rocks (top of subducting plate). The thermal state depends on the age of the incoming plate and the rate of subduction (Stein and Stein, 1996).
- Fluid overpressure along the interface caused by the progressive dehydration of the subducting plate. Fluid pressure at a given depth depends on the available input from the subducting slab, temperature, and pressure that control phase changes and permeability of the rocks of the hanging wall (see review by Saffer and Tobin, 2011).

The great size of the fault and the low temperature increases the seismic potential. This is the reason why all giant earthquakes occur at subduction zones. When the subducting plate is old and the subduction rate is fast (e.g., below Japan), the subducting plate is characterized by a cloud of intraslab earthquakes termed the “Wadati-Benioff” or “Benioff” zone (Wadati, 1928; Benioff, 1949, see entry “▶ Wadati-Benioff Zone”) down to 660 km, which is the depth of the discontinuity between the upper and the lower mantle.

The last distinctive characteristic is the volcanic arc that generally outlines the oceanic subduction zones. Fluids stored in the subducting plate are released at various depths from the first kilometers of subduction near the trench (pore fluids) down to about 150 km (hydrous minerals) depending on the thermal state of the slab. At depths between 80 and 150 km, slab dehydration is responsible for the metasomatism of the overlying mantle wedge. A small fraction of the asthenospheric wedge undergoes partial melting and migrates upward. Volcanoes thus develop at a given distance depending on the slab dip because melts migrate vertically from the slab.

**Slab Shape and Dynamics**

At a large scale, oceanic plates behave elastically when they are forced to bend in subduction zones. We often observe a small bulge seaward of the trench before the plate sinks into the mantle. The trench is the deepest surface expression of the oceanic plate. Its depth depends on the age of the subducting plate and the amount of sediment infill. The deepest trenches are the Mariana Trough (10,938 m revised depth by Fujioka et al., 2002) and the Philippine trench (10,100 m after
Lallemand et al., 1998). It is interesting to note that the deepest trenches are located right in the area of maximum downwelling of oceanic lithosphere on Earth as computed by Lithgow-Bertelloni and Richards (1998). Some authors argue that the elastic properties of the sinking lithosphere help the slab to unbend below the arc so that it recovers a flat profile. In detail, other forces may also contribute in the unbending of the plate such as the suction force exerted by the overriding plate and mantle wedge. Slabs may exhibit various shapes such as flat and horizontal in their upper section (Peru or Central Chile), shallow dipping (Japan or Kurile), intermediate dipping (Tonga or Kermadec), steeply dipping (Izu-Bonin), vertically dipping (Mariana), or even overturned (Luzon or New Britain).

In modern subduction zones, active compression in the overriding plate is always observed for shallow-dipping slabs (<50°), whereas active back-arc extension is always observed for steeply dipping slabs (>50°). Lallemand et al. (2005b, 2008) have shown that the slab shape is not simply linked with the age of the subducting plate (the common assumption is that an old plate is thick and dense and thus prone to sink steeply or even vertically in the mantle) but rather with the combination of absolute velocities between converging plates and the regional mantle dynamics. Moreover, the upper plate strain appears to directly result from this kinematic interaction.

**Main Driving and Resistive Forces in Subduction Zones**

Today, the main driving mechanism for plate motion in general (see the entry “Driving Forces”), and for subduction in particular, is the gravitational sinking of the slab (Turcotte and Schubert, 1982) which is usually slightly denser (about 1 %) than the surrounding mantle because of its lower temperature (thermal contraction). This force is called *slab pull* (Fig. 2). The viscous mantle opposes
a resistance to plate motion from the ridge to the trench called mantle drag. The upper portion of the asthenosphere is known to have a low viscosity so that the mantle drag should be moderate. The viscous shearing increases considerably as the plate sinks into the mantle because it exerts on both sides – top and bottom – of the descending slab and because the viscosity is pressure dependent (Karato and Wu, 1993). It probably increases by at least one order of magnitude between the top and the bottom of the upper mantle. This resistance to penetration drops again when the slab reaches the discontinuity between the upper and the lower mantle near 660 km as a result of phase changes in minerals. The mantle opposes another resistance called anchoring, which accounts for mantle reaction to slab facewise translation (Scholtz and Campos, 1995; Lallemand et al., 2008). Since it is a viscous pressure force, it will depend on the velocity of slab rollback or advance associated with trench migration and increase with depth. The slab also resists to bend and unbend when passing the trench. This resisting force called bending moment increases with the stiffness of the plate and thus its age. It is inversely proportional to the radius of curvature of the slab (Buffett and Rowley, 2006; Wu et al., 2008).

Among these main internal forces, two are always resisting: the mantle resistance to penetration, by definition, which is effective as soon as a slab dives into the mantle, and the bending moment, also called bending resistance, as the plate bends and unbends. All the other forces can shift from driving to resisting depending on the geodynamic context. Let us first consider slab pull whose name comes, by definition, from the capacity of the slab to sink into the mantle because of its negative buoyancy. Most of the time, it is considered as a driving force, but in about 10% of cases, we observe “flat” slabs, i.e., subducting slabs sliding horizontally beneath the upper plate over hundreds of kilometers before sinking. Gutscher et al. (2000) attributed this flatness to a positive buoyancy of the oceanic plate and a delay in the basalt to eclogite transition due to the cool thermal structure of the two overlapping lithospheres. The best examples are the Nankai or Cascadia subduction zone where the subducting crust is very young, or the Peru and Central Chile segments along the Andean margin where buoyant ridges or plateaus are subducting. The mantle drag generally offers resistance to plate motion near subduction zones except when mantle convective cells, produced by another mechanism rather than the subduction, drive the plate down. This could be the case when a spreading ridge subducts like in South Chile or in Cascadia. The anchoring force always offers resistance to facewise translation of the slab but not much to its downward penetration.

There are also second-order external forces originating from the far field (Fig. 2). Their magnitude and direction depend on the geodynamic context. The so-called ridge-push results from the gravitational collapse or subsidence of the oceanic plates as they drift away from the ridge. Fowler (1990) estimates that, on average, the slab pull force is one order of magnitude greater than the ridge-push force. If the active ridge is located seaward of the trench, the ridge-push force will push the plate toward the trench, but if the spreading ridge is located landward of the trench, it will push the upper plate onto the subducting plate (like in South America with the East Pacific Rise on one side and the mid-Atlantic Ridge on the other). Far-field slab pull can also provide an additional force acting in the subduction zone. For example, the Izu-Bonin-Mariana arc overrides the subducting Pacific plate but is carried by the Philippine Sea plate which is dragged northwestward as a consequence of the slab pull acting along the Ryukyu subduction zone (Pacanovsky et al., 1999). In this case, the upper plate is pulled away from the trench. Another source of external stress is the effect of a collision like the Ontong Java Plateau with the Solomon trench that impacts the stress field over a huge area from New Guinea to New Hebrides (Mann and Taira, 2004).
**Subduction Interface**

The shallowest part of the subduction interface often corresponds to an aseismic décollement zone characterized by overpressures resulting from the pore fluids release as lithostatic pressure increases (Saffer and Tobin, 2011). At rather low temperatures between 100–150°C and 350–450°C (Oleskevich et al., 1999; Hyndman, 2007), the plate interface becomes frictional and thus seismogenic (see entry “Seismogenic Zone” and Dixon and Moore, 2007). All “subduction earthquakes” nucleate in this region. The seismogenic zone generally extends from a depth of 11 ± 4 to 51 ± 9 km. It is about 112 ± 40 km wide and dips 23 ± 8° (Heuret et al., 2011). At temperatures higher than 350–450°C, the slip along the interface is either continuous (creep) or characterized by slow-slip events (SSE) associated with nonvolcanic tremors (NVT) (Rogers and Draggert, 2003). The downdip limit of the subduction interface is marked by the coupling between the descending plate and the overlying mantle (Fig. 3). Furukawa (1993) called the depth below which the plates are uncoupled on the long term $Z_{\text{dec}}$. Downdip of this depth, a significant thickness of mantle is dragged with the slab, called “viscous blanket” by Kincaid and Sacks (1997). The thickness depends on the temperature contrast between the slab and the mantle wedge. Beyond $Z_{\text{dec}}$, there is no more plate interface, the slab being overlain by the convective mantle wedge.

There is a debate on the nature of the plate interface. Is it an ~1–20 m thin “décollement” zone like those drilled in Nankai (Bangs et al., 1996) or an ~100–1,000 m thick channel across which the shear distributes like in the exhumed Franciscan complex (Cloos and Shreve, 1988)? The discovery of erosional processes at the hanging wall of the subduction interface in Japan, Middle America, and Izu-Bonin-Mariana trenches since the early 1980s (von Huene et al., 1980; von Huene and Lallemand, 1990; von Huene and Scholl, 1991; Lallemand et al., 1992; Meschede et al., 1999) has been decisive in promoting the concept of subduction channel. It has been later confirmed,

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**Fig. 3** Schematic illustration of the subduction interface in an oceanic subduction zone. $U_z$ is the upper limit and $D_z$ the downdip limit of the seismogenic zone. $Z_{\text{dec}}$ is the decoupling depth. Below that depth, the convective mantle is dragged down (coupled) with the subducting plate. NVT nonvolcanic tremors and SSE slow-slip events are observed in some subduction zones at depths between $D_z$ and $Z_{\text{dec}}$. 

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thanks to the increasing resolution of seismic images as in the Ecuadorian margin (Sage et al., 2006; Collot et al., 2008) and the reinterpretation of some outcrops of exhumed subduction interface like in the Shimanto belt (Kitamura et al., 2005) or in the Apennines (Vannucchi et al., 2008). Several processes of upward migration of the roof décollement or downward migration of the basal décollement delimiting the subduction channel have been discussed by Vannucchi et al. (2012) and are not yet completely understood. Hydrofracturing of the hanging wall rocks under increasing fluid pressure coming from the dewatered downgoing plate (von Huene et al., 2004) very probably contributes to the upward migration of the deformation front as well as the subduction of oceanic reliefs such as seamounts and ridges (Dominguez et al., 1998, 2000).

At the scale of the plate interface, two forces govern local tectonics and hazards: friction and pressure. The net pressure or normal stress exerted along the subduction interface combines the lithostatic pressure, increasing with depth, and the non-isostatic “dynamic” pressure, which accounts for the slab pull, the slab-mantle interactions, and the far-field “tectonic” forces. The idea, developed by Shemenda (1992) in laboratory analog models, where far-field forces and mantle drag and anchoring are neglected, is that a dense (old) and long subducting plate tends to pull the upper plate down producing a subsidence of the forearc, whereas a light (young) or a short subducting plate tends to push the upper plate up producing uplift of the forearc. In nature, not only the age and length of the subducting plate but mantle and far-field forces also contribute to the normal stress along the interface. The second force is frictional shear, which resists subduction. The stress is released intermittently as it overcomes the yield strength of the fault, through repeated earthquakes and subsequent fault ruptures.

**Seismicity with Special Emphasis on Mega-thrust Earthquakes**

Subduction zones are the locus of most earthquakes occurring on Earth because the plate boundaries have to accommodate the largest relative motions and subsequent strain, because the plate interface is continuously cooled and because the slab still remains cool at large depths. “Subduction earthquakes” are thrust events occurring along the seismogenic plate interface. They represent most of the seismic energy released at subduction zones and in the world (Scholz, 2002). There are also intraslab shallow, intermediate, and deep earthquakes that accommodate the high level of stress during the plate bending and penetration into the viscous mantle and the upper plate earthquakes that accommodate the part of the stress transmitted through the plate interface. See entries “▶ Earthquake”, “▶ Active Continental Margin” and in particular Fig. 2 of “▶ Active Continental Margin.”

One specificity of the subduction zones is that they can generate giant earthquakes with magnitudes \( \geq 8.5 \), most of them being tsunamigenic like the 2004 M9.2 event off Sumatra (Indonesia) or the 2011 M9.0 event off Tohoku (Japan). Back to the 1900–2012 period, Fig. 4 shows the rupture areas associated with the M \( \geq 8.0 \) earthquakes and the segmentation of the seismogenic zones mainly based on past earthquake ruptures. It is easy to see that they don’t distribute equally along the subduction zones. Based on available global datasets, Heuret et al. (2011, 2012) have shown that giant earthquakes preferentially occurred near slab edges, where the upper plate was continental, the back-arc strain neutral to compressive, and the trench fill larger than 1 km. A possible explanation would be that the rupture propagation could be facilitated if the plate interface is smooth (Ruff, 1989) and the normal stress on the interface large enough to accumulate strain but not too large to allow rupture propagation. The continental nature of the upper plate could be preferred because it is associated with wider (downdip) seismogenic zones. The dependence of earthquake magnitude with
Fig. 4 Map of the subduction interface seismic rate, trench segments, and major rupture areas of the $M_w \geq 8.0$ 1900–2012 events updated from Heuret et al. (2011). Pale blue arrows indicate subduction relative velocities after Heuret and Lallemand (2005).
the subduction velocity and plate age, first proposed by Ruff and Kanamori (1980, 1983), is not statistically verified using revised databases. However, none of the Sumatra and Tohoku earthquakes were expected to occur based on our knowledge at the time they happened, because the recurrence time of such huge events is often underestimated and because we generally do not have access to paleoearthquake archives (McCaffrey, 1997, 2008; Lay and Kanamori, 2011).

The degree of coupling between the converging plates may be used to estimate the seismic hazard at a given subduction zone. For many years, authors calculated the seismic coupling coefficient as the ratio between the observed seismic moment release rate and the rate calculated from plate tectonic velocities (Brune, 1968; Davies and Brune, 1971). Scholz and Campos (1995) have shown that this coefficient should be used cautiously because the seismic slip was generally computed over a short time period (typically a hundred years) compared with the recurrence time for large earthquakes (sometimes up to a thousand years). During the last two decades, the advent of monitoring of subduction zones with GPS stations allowed geodesists to measure the interseismic strain accumulation rate in the upper plate, which gives a better picture of the interplate coupling than the earlier measure of energy release rate by summing the earthquake moment over a short period. By doing this, Scholz and Campos (2012) have revised their estimate of the seismic coupling coefficient in many subduction settings. They concluded that this new approach agrees well with the revised “classical way” of measuring the coefficient accounting for major earthquakes older than a century when available. It is also more satisfactory for the balance of forces acting along the plate interface like the normal stress and the friction making assumptions on their respective origins (see above). They confirm the observation of Chemenda et al. (2000) that high coupling, and thus high friction, is observed in regions of high normal stress. The limitation of the method, consisting of measuring the interseismic elastic strain, is that most of the time the coupled area in a subduction zone is located offshore, whereas GPS receivers are deployed onshore. The resolution of the models should be improved in the future when a significant number of GPS stations can be installed offshore like off Japan. Tremendous developments must be further done to precisely map the complexity of the coupling distribution along the subduction interface because large lateral variations are predicted by authors (Ruff, 1992; Bilek, 2007; Scholz and Campos, 2012).

**Upper Plate Tectonics**

The upper plate is the visible part of a subduction zone, also called active margin. The tectonic regime of the active margin depends on interactions between the converging plates at the level of the trench, the plate interface, and even the whole overriding plate.

At the trench, sediment carried by the subducting oceanic plate may be offscraped and incorporated at the front of the overriding plate. This process, called frontal accretion, contributes to the growth of an accretionary wedge (see entry “Accretionary Wedge” for more details or Moore and Silver, 1987). Von Huene and Scholl (1991) have shown that half of the growth of an accretionary prism can be accounted for by subcrustal or basal accretion, also called underplating. They also estimated that only half of the modern convergent margins are accretionary. The other half undergoes what they called tectonic erosion. This process has been definitely demonstrated in the late 1970s (e.g., Scholl et al., 1980) after observations from both deep-sea drilling and reflection seismic imaging, which the Japan margin underwent a Miocene subsidence of at least 2 km. Such large subsidence increasing seaward, together with the absence of an accretionary prism, erosional features at margin’s front, and simultaneous retreat of the volcanic arc, led the authors to propose a model in which part of the margin was removed both frontally and subcrustally during subduction.
Since 1980, other margins appeared to also undergo tectonic erosion like in mid-America, Peru, Tonga, Mariana, or Izu-Bonin (von Huene and Lallemand, 1990; Lallemand, 1995; Meschede et al., 1999; Clift and Vannucchi, 2004). See entry "Subduction Erosion" for more details.

The convergent margins deform in response to the stress transmitted across the plate interface but also in response either to frontal or basal accretion or frontal or basal tectonic erosion. The tectonic regime is often compressive in the frontal part of the overriding plate, but it is also quite common to observe normal faults dipping either landward or seaward in active convergent margins, especially when they are subject to tectonic erosion. These faults reflect either a trenchward collapse as the margin progressively steepens (Aubouin et al., 1984), or a buckling as underplating occurs (Lallemand et al., 1994). Another mechanism, coseismically induced, has been recently suggested. Indeed, a seaward dipping normal fault has been activated at the front of the Japan margin during the 2011 Tohoku mega-earthquake. The explanation provided by McKenzie and Jackson (2012) would be that both the release of gravitational potential energy and the elastic strain account for this coseismic behavior.

Trench-parallel strike-slip faults may also be observed in the forearc in case of oblique convergence like off Sumatra (Mentawai Fault) or the southern Ryukyus (Yaeyama Fault). These transcurrent subvertical faults accommodate part of the lateral component of the oblique convergence, whereas the slip along the frontal part of the subduction thrust is closer to trench normal (e.g., McCaffrey, 1992). This mechanism is verified by the direction of slip vectors of subduction earthquakes that are often deflected toward trench normal when the convergence is oblique. This process called slip partitioning has been observed not only at the scale of the forearc but also at the scale of tectonic plates. Indeed, several oblique subduction systems develop transcurrent faults. Chemenda et al. (2000), for example, have investigated using physical models the effect of the pressure and the friction along the plate interface on slip partitioning in a context of oblique convergence. They conclude that slip partitioning can occur in the models only when interplate friction is high and when the overriding plate contains a weak zone. As a matter of fact, lithospheric transcurrent faults develop either at the toe of the rigid backstop (rear of accretionary wedge like off East Taiwan), along the volcanic arc (Sumatra, the Philippines, South Kuriles), at the back-arc rift axis, or spreading center (Andaman Sea, Le Havre Trough).

Ultimately, the global strain rate map (Kreemer et al., 2003) indicates that most subduction systems deform far from the plate boundaries, or one may consider that the plate boundaries are wider than a single fault. They can potentially affect a strip several hundreds of kilometers large at upper plate edge. In addition to interseismic elastic deformation, there are many areas where permanent (plastic) deformation localizes. We have seen above that, in some oblique geodynamic settings, transcurrent faulting may occur, but even more shortening, rifting, or spreading can also occur. Significant shortening (>1 cm/year) is observed, for example, along the eastern margin of the Japan Sea, the southern margin of the Okhotsk Sea, or the eastern cordillera of the Andes. Oppositely, rifting or spreading develops in the Sumisu rift and the Mariana Trough (Izu-Bonin-Mariana arc), the Lau Basin and Le Havre Trough (Tonga-Kermadec arc), the Andaman Sea (Andaman arc), the Central Kamchatka graben (North Kurile arc), the Manus Basin (New Britain arc), the North and South Fiji basins (New Hebrides arc), the Scotia Sea (South Sandwich arc), the Tyrrhenian Sea (Calabria arc), or the Aegean Sea (Hellenic arc). Several mechanisms have been invoked to account for this upper plate deformation playing with interplate stress resulting from plate kinematics (Dewey, 1980; Lallemand et al., 2008) and/or mantle wedge dynamics (Sleep and Toksöz, 1971; Lagabrielle et al., 1997; Faccenna et al., 2010).
Magmatism at Subduction Zones

Volcanic arcs are the second worldwide magmatic contributor to crustal growth on Earth after the oceanic spreading centers (nearly 1 km³/year according to Crisp, 1984). Most subduction zones are characterized by an active volcanic arc. Sometimes, the volcanic activity ceases for a few million years and resumes again. The reason for the cessation of volcanism is commonly either the collision with a continent or the geometry of the slab which is incompatible with the presence of a mantle wedge. This is the case, for example, with flat slab segments in the Andes (Gutscher et al., 2000). Arc magmatism results from the metasomatism and partial melting of rocks of the mantle wedge that are hydrated by the water extracted from hydrous minerals in the subducting plate at depths ranging between 80 and 150 km. Further details on arc magmatism can be found in Tatsumi (1989) or Tatsumi and Kosigo (2003). See also entries “Magmatism at Convergent Plate Boundaries”, and “Island Arc Volcanism, Volcanic Arcs”.

Access to a database on kinematic, geometric, seismologic, and structural parameters of oceanic subduction zones can be found via url: submap.fr.

Summary

Subduction is a lithospheric process which occurs at convergent plate boundaries. One plate underthrusts another and sinks into the Earth mantle. The subduction mainly involves a subducting oceanic plate, but sometimes, a continental plate enters into the subduction zone. In this last case, a collisional chain develops. This process generates a high level of seismicity as the rocks deform in the vicinity of the plate boundary. All mega-earthquakes occur in subduction zones. Oceanic subduction creates favorable conditions for arc magmatism.

Cross-References

- Accretionary Wedge
- Active Continental Margin
- Crustal Accretion
- Driving Forces
- Earthquake
- Intraoceanic Subduction Zone
- Island Arc Volcanism, Volcanic Arcs
- Lithosphere-Composition and Formation
- Magmatism at Convergent Plate Boundaries
- Marine Evaporites
- Morphology Across Convergent Plate Boundaries
- Ocean Margin Systems
- Orogeny
- Seamounts
- Seismogenic Zone
- Subduction Erosion
- Wadati-Benioff Zone
- Wilson Cycle-Marine Geosciences
Bibliography


