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# Convergent plate margin dynamics: New perspectives from structural geology, geophysics and geodynamic modelling

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ABSTRACT

Convergent plate margins occur when two adjoining tectonic plates come together to form either a subduction zone, where at least one of the converging plates is oceanic and plunges beneath the other into the mantle, or a collision zone, where two continents or a continent and a magmatic arc collide. Convergent plate margins are arguably the most complicated and dynamic plate boundaries on Earth and have been the subject of many investigations and discussions since the advent of plate tectonic theory. This paper provides a historical background and a review of the development of geological and geodynamic theories on convergent plate margins. Furthermore, it discusses some of the recent advances that have been made in the fields of structural geology, geophysics and geodynamics, which are fundamental to our understanding of this phenomenon. These include: (1) the finding that plates and plate boundaries move at comparable velocities across the globe; (2) the emerging consensus that subducted slabs are between two to three orders of magnitude stronger than the ambient upper mantle; (3) the importance of lateral slab edges, slab tearing and toroidal mantle flow patterns for the evolution of subduction zones; and (4) clear evidence from mantle tomography that slabs can penetrate into the lower mantle. Still, many first-order problems regarding the geodynamic processes that operate at convergent margins remain to be solved. These include subduction zone initiation and the time of inception of plate tectonics, and with it convergent plate margins, on Earth. Fundamental problems in orogenesis include the mechanism that initiated Andean mountain building at the South American subduction zone, the potential episodicity of mountain building with multiple cycles of shortening and extension, and the principal driving force behind the construction of massive mountain belts such as the Himalayas–Tibet and the Andes. Fundamental questions in subduction dynamics regard the partitioning of subduction into a trench and plate component, and the distribution of energy dissipation in the system. In seismic imaging, challenges include improving resolution at mid to lower mantle depth in order to properly understand the fate of slabs, and better constraining the 3-D flow-related anisotropic structure in the surrounding mantle. Future insights into such fundamental problems and into the regional and global dynamics of convergent plate margins will likely be obtained from integrating spatio-temporal data, structural geological data, geophysical data and plate kinematic data into plate tectonic reconstructions and three-dimensional geodynamic models of progressive deformation.



## 1. Introduction

Convergent plate margins have traditionally been subdivided into two categories: subduction zones and collision zones (Fig. 1). At subduction zones one plate (the subducting or lower plate) sinks (subducts) below the other plate (overriding plate or upper plate) into the mantle. The subducting plate is an oceanic plate, whilst the overriding plate can be either oceanic or continental (Fig. 1A–B). At collision zones, both plates are continental in nature (Fig. 1C–D), or one is continental and the other carries a magmatic arc.

Although the simple subdivision of convergent plate margins is sensible, a quick glance at the active convergent margins on Earth makes it clear that there is a large variability in structure of both subduction zones and collision zones. As such, subduction zones might contain elements that are more typical of collision zones and vice versa. For example, a collision zone is characterized by a large mountain range, but the Andes, the longest mountain range on Earth (>7000 km), is located at the edge of the South American subduction zone, and different theories as to why a major mountain belt is located at a subduction zone abound (e.g. Molnar and Atwater, 1978; Uyeda and Kanamori, 1979; Jarrard, 1986; Russo and Silver, 1996; Somoza, 1998; Silver et al., 1998; Gutscher et al., 2000; Lamb and Davis, 2003; Oncken et al., 2007; Schellart, 2008a). Another example is the Himalayas–Tibet mountain belt, which formed since ~50 Ma due to

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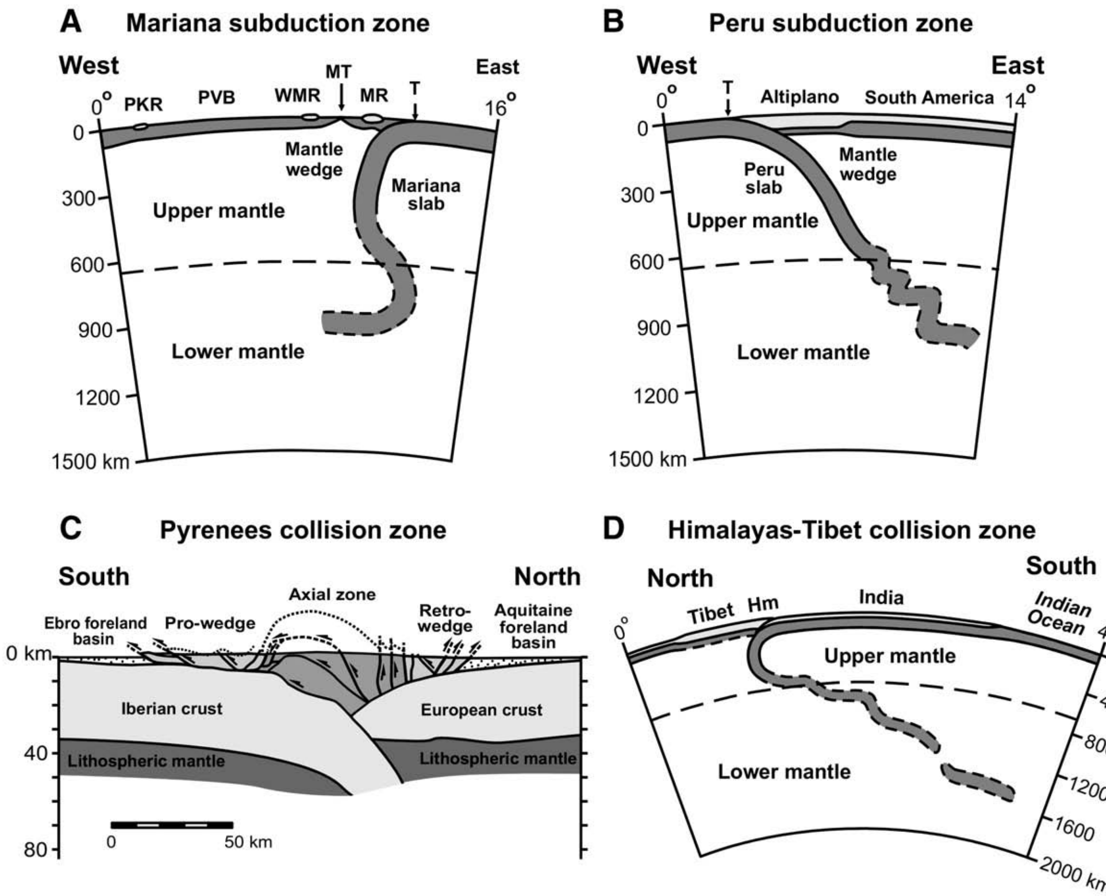


**Fig. 1.** Diagrams showing the two types of convergent plate margins, namely subduction zones (A and B) and collision zones (C and D). (A) The Mariana subduction zone as an example of an ocean–ocean subduction zone with an oceanic overriding (upper) plate (Philippine plate) and an oceanic subducting (lower) plate (Pacific plate). Modified from Schellart (2005) with slab structure interpreted from tomography in Widiyantoro et al. (1999). Note the active Mariana Ridge volcanic arc (MR), the remnant volcanic arcs (WMR–West Mariana Ridge, PKR–Palau–Kyushu Ridge), the active Mariana Trough backarc basin (MT) and the inactive Parece–Vela backarc basin (PVB). (B) The Peru subduction zone as an example of an ocean–continent subduction zone with a continental overriding (upper) plate (South American plate) and an oceanic subducting (lower) plate (Nazca plate). Slab structure interpreted from tomography in Li et al. (2008b). Note the thickened crust (up to ~70 km) in the Altiplano region. (C) The Pyrenees mountain belt as an example of a continent–continent collision zone, with two converging continental plates (Iberian plate and Eurasian plate). Simplified from Schellart (2002) but originally from Muñoz (1992). Note that the Iberian plate is the underthrusting plate. (D) The Himalayas–Tibet mountain belt as an example of a continent–continent collision zone between the Indian plate and the Eurasian plate with apparent penetration of Indian lithosphere into the sub-lithospheric mantle down to >600 km. Modified from Schellart (2005) with slab structure interpreted from tomography in Van der Voo et al. (1999). Note the thickened crust (up to ~80 km) in the Tibet region. Hm—Himalayas.

collision of two continents: India and Eurasia (e.g. Argand, 1924; Dewey and Bird, 1970; Molnar and Tapponnier, 1975; Searle et al., 1987). Seismic tomography studies of this collision zone reveal local high-velocity anomalies that could represent continental Indian lithosphere segments dipping steeply below the Himalayas down to several hundred kilometres depth or more (Fig. 1D) (e.g. Van der Voo et al., 1999; Replumaz et al., 2004; Li et al., 2008a; Replumaz et al., 2010-this issue). Note, however, that individual tomography models show discrepancies in horizontal extent, vertical extent and geometry of these high-velocity anomalies in the mantle below this collision zone, and there is also a conspicuous absence of a continuous planar Wadati–Benioff zone.

In its final phase of existence, a subduction zone may become a collision zone once the entire ocean basin in between the convergent plates has been consumed (Dewey and Bird, 1970). This is generally thought to have occurred along most of the Alpine–Himalayan chain, when the Tethys Ocean and smaller marginal basins were consumed during the accretion of arc terranes and continental ribbons, and when the Adriatic promontory, the Arabian continent and the Indian continent collided with Eurasia. Seismic tomography shows ample evidence for the consumption of these Tethyan ocean basins in the form of high-velocity anomalies in the upper and lower mantle below and south of the Alpine–Himalayan chain (e.g. Bijwaard et al., 1998; Van der Voo et al., 1999; Replumaz et al., 2004; Hafkenscheid et al., 2006).

At present, the globe is covered with numerous active convergent margins in the form of (mature) subduction zones, incipient subduction zones, continental collision zones and arc–continent collision zones (Fig. 2). Most subduction zones are found along the margins of the Pacific Ocean, whilst some are found in the Indian Ocean, the Caribbean, the Mediterranean and the southern Atlantic. The total length of these subduction zones amounts to some 48,800 km, whilst the total length of incipient subduction zones amounts to some 10,550 km (Schellart et al., 2007). Most collision zones are found along the Alpine–Himalayan chain and the total length of these collision zones amounts to some 23,000 km (Bird, 2003). This results in a total of 82,350 km of convergent plate boundaries on Earth.

This paper presents a historical outline and review of the development of geological theories of large-scale tectonic processes, in particular focussing on convergent plate boundaries. It will thereby show that contributions from structural geologists, geophysicists and geodynamic modellers have been crucial for the development of these theories and for the understanding of convergent plate boundaries.

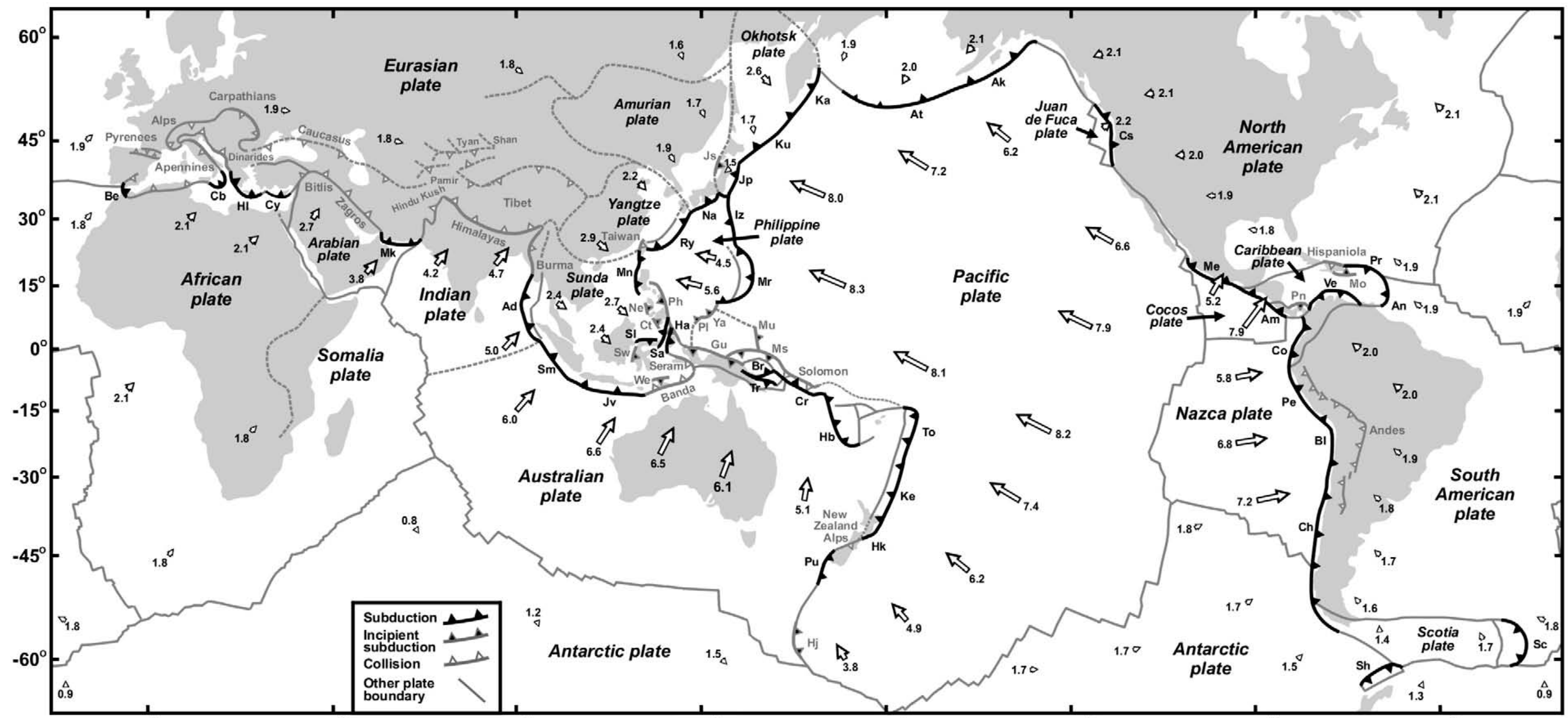


**Fig. 2.** Global tectonic map showing all subduction zones on Earth (including incipient subduction zones), collision zones, and the velocities of the major plates in the Indo-Atlantic hotspot reference frame from O'Neill et al. (2005) using the relative plate motion model from DeMets et al. (1994). Subduction zones: Ad—Andaman, Ak—Alaska, Am—Central America, An—Lesser Antilles, At—Aleutian, Be—Betic–Rif, Bl—Bolivia, Br—New Britain, Cb—Calabria, Ch—Chile, Co—Colombia, Cr—San Cristobal, Cs—Cascadia, Cy—Cyprus, Ha—Halmahera, Hb—New Hebrides, Hk—Hikurangi, Hl—Hellenic, Iz—Izu–Bonin, Jp—Japan, Jv—Java, Ka—Kamchatka, Ke—Kermadec, Ku—Kuril, Me—Mexico, Mk—Makran, Mn—Manila, Mr—Mariana, Na—Nankai, Pe—Peru, Pr—Puerto Rico, Pu—Puysegur, Ry—Ryukyu, Sa—Sangihe, Sc—Scotia, Sh—South Shetland, Sl—North Sulawesi, Sm—Sumatra, To—Tonga, Tr—Trobriand, Ve—Venezuela. Incipient subduction zones: Ct—Cotobato, Gu—New Guinea, Hj—Hjort, Js—Japan Sea, Mo—Muertos, Ms—Manus, Mu—Mussau, Ne—Negros, Ph—Philippine, Pl—Palau, Pn—Panama, Sw—West Sulawesi, We—Wetar, Ya—Yap.

## 2. The first geological theories

### 2.1. The first ideas of a deforming Earth

The mobilist view of the Earth, which culminated in the theory of plate tectonics in the 1960s, was preceded by a long history of geological data collection and development of geological theories. Probably the most important advance in understanding the Earth was the appreciation by mankind that the Earth is spherical rather than flat, a fact that became generally accepted in the 16th century. Also in the 16th century, cartography of the Earth's landmasses commenced thanks largely to exploration by the Spanish, Portuguese, Dutch, English and French. Since the first compilations of global geographical maps people have been intrigued by the similarity of the coastlines of Africa and Europe on one side and the Americas on the other. Possibly the first person who noticed this similarity and proposed an ancient separation between these continents was the Dutch cartographer Abraham Ortelius in 1596 (Rom, 1994). This hypothesis implies a mobilistic view of the Earth with large horizontal displacements between the different landmasses of the order of thousands of kilometres.

In that same century, theories were developed to explain the occurrence of mountain belts. One theory proposed to explain mountain formation was the contracting Earth theory due to planetary cooling, for which the Italian philosopher Giordano Bruno (1548–1600) was one of the early founders (Dennis, 1982). This theory was later expanded by the French philosopher Descartes (1644) in his book “Principia Philosophiae”. Much later, in the 18th century, the Scotsman James Hutton (Hutton, 1788) proposed that thermal expansion of the Earth caused the formation of elevations including mountain belts, continental platforms, and most folding of rocks.

### 2.2. Large-scale horizontal motions

An important insight into mountain building came from De Sausurre (1784–1796, 1796), who proposed that lateral forces were the cause for shortening, tilting and folding of originally horizontal strata (Dennis, 1982). The original ideas of De Sausurre (1784–1796, 1796) gained more ground in the 19th century as it was convincingly demonstrated that the internal structure of the Alps had resulted from large-scale horizontal shortening, involving northward overthrusting of the whole range (Suess, 1875). At the same time, these ideas were experimentally investigated with analogue models (small-scale physical laboratory experiments), which provided valuable new insights into structures of the Earth. The results verified that folding and thrusting of rocks, as well as mountain formation, resulted from horizontal compression rather than vertical compression (Hall, 1815; Favre, 1878; Daubre, 1879; Schardt, 1884; Cadell, 1889; Willis, 1893). Suess continued to explain this shortening within the context of the contracting Earth theory (Suess, 1875, 1885).

Near the end of the 19th century and the start of the 20th century, geologists started to have doubts about the contraction theory as a mechanism for orogenesis (Dennis, 1982). Fisher (1881) expressed his doubts about the ability of a cooling and contracting Earth to produce large amounts of crustal shortening as observed in the Alps. Dutton (1889) raised further doubt by reasoning that a contracting Earth would result in shortening of the Earth's crust in each direction equally and would not explain the occurrence of long narrow belts of parallel folds.

## 3. Theory of continental drift

In 1912 Alfred Wegener proposed his theory of continental drift (Wegener, 1912). Wegener put forward a large amount of data from various disciplines to support the hypothesis that North and South America were once connected to Europe and Africa and formed a supercontinent, which he called Pangea. The continents were later separated from each other due to divergent motion between these landmasses. Wegener compiled much of the pre-drift geological data to show that the continuity of older structures, formations and fossil floras and faunas located along the shorelines of many continents could be explained on a pre-drift reconstruction of Pangea. The geoscience community received the continental drift theory with disbelief, since the theory broke with established orthodoxy of a static Earth. Furthermore, many geoscientists found faults in the details of the geological and geophysical data and interpretations outlined in Wegener's work. Finally, most of the geophysical community at that time rejected the theory, since Wegener, as he himself realised, did not propose a reasonable mechanism for continental movements. Wegener had suggested that continental drift was powered by the centripetal force, which would cause high standing continents to move due to rotation of the Earth. However, it was soon realised that this force was much too small to drive continental movement.

In 1928 Holmes postulated that continental drift could be driven by convection currents from deep within the Earth, which in turn were powered by the heat in the interior of the Earth generated by radioactive decay (Holmes, 1928). The idea of sub-crustal convection in the Earth's mantle was not new and had already been proposed a century earlier (Hopkins, 1839; Fisher, 1881). In the beginning of the twentieth century it was proposed by several geoscientists that these mantle convection currents could also drive the formation of mountain belts and their associated fold and thrust structures (Ampferer, 1906; Schwinner, 1920; Vening Meinesz, 1934). These theories were later supported by results from analogue experiments (Griggs, 1939), which showed that convection in a fluid substratum can cause shortening in an overlying plastic crust.

## 4. Theory of plate tectonics

After the publication of Wegener's drift theory (Wegener, 1912, 1915), the geoscience community was divided into two schools of thought: those in favour of continental drift and those against continental drift. In the thirty years or so following the publication of Holmes' convection model, very little was written about the continental drift theory. However, this was about to change when in the 1950s and 1960s new geological and geophysical techniques were developed which produced a wealth of new data. These included radiometric dating, bathymetric mapping of the seafloor, the discovery of magnetic reversals patterns and oceanic spreading centres, paleomagnetic data and magnetic polar wandering. The interpretation of these new data sets in the 1960s resulted in a revolution in geology, which led to the development of the theory of plate tectonics.

To begin with, in the early 1960s, Harry Hess and Robert Dietz (Dietz, 1961; Hess, 1962) proposed the concept of seafloor spreading, which stated that seafloor comes into being at mid-oceanic ridges to fill the cracks or voids that are being created by the surrounding seafloor that is moving away in opposite directions on either side of the ocean ridge. This model provided a viable mechanism for the supposed continental drift of continents as proposed by Wegener (Wegener, 1912, 1915), where the continents were split up and pushed apart during the formation of new oceanic crust instead of continents moving through the oceans.

This theory elegantly explained the relatively young age of the ocean basins, which were found to be no more than ~200 Ma, as well as the observation that the age of the ocean floor increases away from mid-oceanic ridges. The theory also provided explanations for various geological observations such as the topography of the ocean floor (Menard, 1965), normal faulting on the top of mid-oceanic ridges (Ewing and Heezen, 1956) and linear magnetic anomalies on the seafloor (Vine and Matthews, 1963). The magnetic anomalies were discovered in the 1950s and were shown to be remarkably linear and continuous, running roughly parallel to the mid-oceanic ridges. Vine and Matthews suggested that these magnetic lineations might be explained in terms of seafloor spreading and paleomagnetism, where the oceanic crust that is continuously formed along mid-oceanic ridges records the reversal history of the Earth's magnetic field. The theory of seafloor spreading would require either expansion of the Earth or destruction of ocean floor away from ridges. It was soon realised that any expansion hypothesis would require an extremely high rate of expansion of the Earth since the Jurassic, since no ocean floor on Earth is older than Jurassic. Thus, it was more logical to pursue the hypothesis of seafloor destruction.

It was already proposed much earlier by Vening Meinesz that trenches were the sites of seafloor destruction, based on gravity measurements along deep sea trenches in the East Indies (Vening Meinesz, 1934, 1962). This idea was further supported by small-scale analogue models (Kuenen, 1936), in which lateral compression of a plastic layer overlying a viscous substratum initially caused down-buckling of the plastic layer and finally resulted in some sort of primitive subduction in a more advanced stage.

The most convincing evidence for destruction of seafloor came from earthquake seismology, where it was shown that ocean floor is thrust underneath continents or island arcs along trenches (Plafker, 1965; Sykes, 1966). Prior to these observations from Plafker and Sykes, deep seismicity such as that observed along the boundaries of the Pacific by Wadati (1935) and Benioff (1949, 1954) was interpreted to occur on a megafault extending from the Earth's surface down to ~700 km depth. Only since the systematic investigation of intermediate and deep earthquake focal mechanisms by Isacks and Molnar (1969) were these zones properly interpreted as reflecting the internal deformation of sinking lithospheric slabs. From then onwards Wadati–Benioff zones were interpreted as representing the presence of subducted slabs.

In the mid 1960s Wilson (1965) stated the basic assumptions of plate tectonic theory, realising the importance of rigidity of rocks in aseismic areas. Furthermore, Wilson recognized a new class of faults named transform faults, which connect linear belts of tectonic activity and along which motion is parallel to the fault trace and horizontal. The geometrical basis of the theory of plate tectonics was soon established by McKenzie and Parker (1967), Morgan (1968) and Le Pichon (1968). By this time, the Earth was viewed as a mosaic of six major and several minor lithospheric plates in relative motion to each other and bound by divergent, convergent or transform plate boundaries. The geometrical outline of these plates was confirmed by earthquake seismology work of Isacks et al. (1968) showing that the majority of seismic activity occurs along the boundaries of these plates (see Fig. 3 for a typical distribution of large earthquakes across the globe).

In addition to earthquake location, seismic records contain information on the nature of the source radiation pattern and hence the source mechanism. During the establishment of plate tectonic theory in the 1960s, the calculation of earthquake focal mechanisms via teleseismic first motion studies played a crucial role in understanding the nature of plate boundaries (e.g. Stauder and Bollinger, 1966; Sykes, 1967; McKenzie, 1970). Sykes (1967) used data from the then recently established WWSSN (World Wide Standardised Seismograph Network) to show that earthquakes located on the fracture zones that intersect the crests of mid-ocean ridges are characterized by a predominance of strike-slip motion. One of the first studies to compute focal mechanisms for earthquakes occurring beneath island arcs was that by Isacks et al. (1969). They looked at shallow, intermediate and deep-focus earthquakes in the Fiji–Tonga–Kermadec region, and concluded that shallow earthquakes occur between the downgoing slab and overriding plate, and deep-focus earthquakes occur within the downgoing slab as a result of internal compressive stress.

With the increasing size of regional and global seismic networks, and the improvements in data quality and availability, a more refined understanding of subduction zone earthquakes soon followed. For instance, it was recognized quite early on that some subduction zones produced much larger earthquakes than others, which was interpreted in terms of the difference in the mechanical coupling at the interface between the two plates (Kanamori, 1971; Uyeda and Kanamori, 1979). Uyeda and Kanamori (1979) introduced the terms "Chilean-type" and "Mariana-type" to describe strongly coupled and weakly coupled subduction zones respectively. Using precise hypocenter determinations, Hasegawa et al. (1978) revealed that the Wadati–Benioff zone beneath the northeast Japan arc comprises two distinct planes separated by some 30–40 km. Focal mechanisms indicate down dip compression in the upper plane, and down dip extension in the lower plane, a phenomenon that has several different explanations including slab unbending. These double planes are characteristic of many but not all subduction zones.

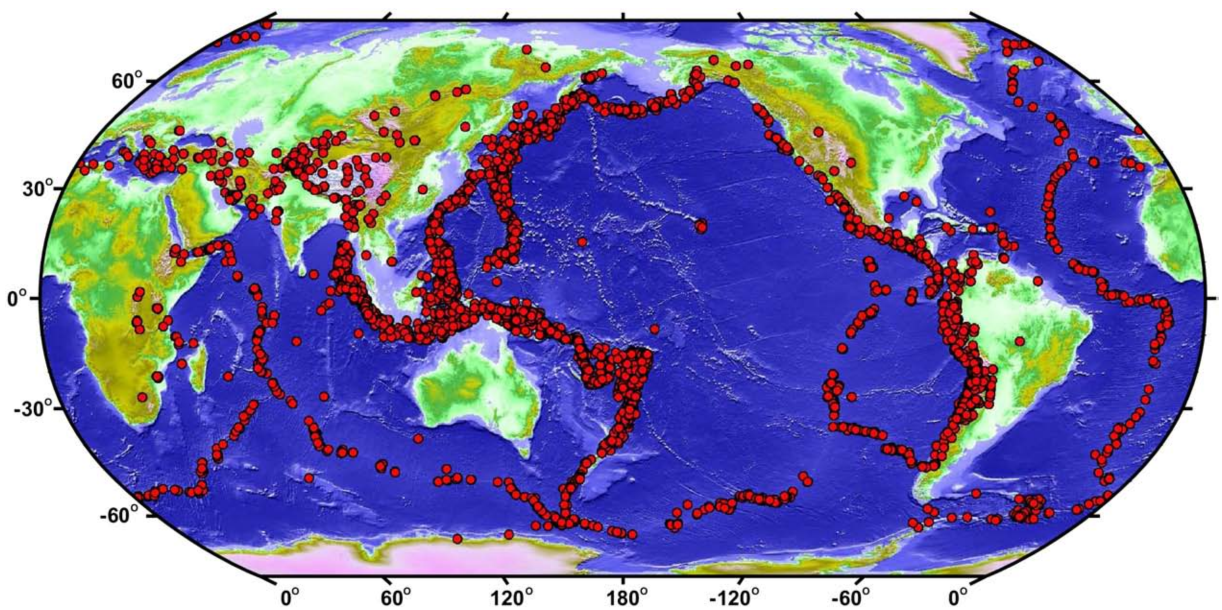


**Fig. 3.** Global map showing the locations of all earthquakes of magnitude 5.0 or greater during the period 01/01/2005 to 01/01/2008. Most large events are concentrated at the boundaries of tectonic plates.

Although the accurate location and characterization of earthquakes has played a major role in the detection and understanding of subduction zones, 2-D and 3-D imaging using seismic tomography has revealed a wealth of additional information. Local earthquake tomography, which exploits seismicity associated with the subduction zone itself, is one of the most popular techniques for high resolution imaging beneath convergent margins. Teleseismic tomography, which makes use of recordings from very distant earthquakes, and wide angle tomography, which relies on active sources, are also common. Early studies include those by Hirahara (1977) and Hirahara and Mikumo (1980) in the Japan region, Spencer and Gubbins (1980) in New Zealand and Spencer and Engdahl (1983) in the Aleutians. Using the techniques developed in the pioneering studies on seismic tomography by Aki and Lee (1976) and Aki et al. (1977), Hirahara and Mikumo (1980) was able to image the structure of the Pacific slab beneath the Sea of Japan to a depth of approximately 600 km. Larger datasets and increasing computing power soon allowed regional and global seismic models to be generated with sufficient detail to detect subduction zones (e.g. Spakman et al., 1988; Inoue et al., 1990; van der Hilst et al., 1991). These larger scale experiments showed that some slabs were perfectly capable of subducting into the lower mantle, a result, which had major implications for mantle convection models.

From several hundred years of observations and geological interpretations of the Earth, a general acceptance of the theory of plate tectonics was achieved in the late 1960s. However, it was soon recognized that although the theory worked well for describing the structures and movements of the oceanic plates, it did not work as well for the continental lithosphere (Isacks et al., 1968) (see also Fig. 3). As observed from the distribution of shallow seismicity on Earth, deformation in oceanic lithosphere is observed along narrow strips up to several tens of kilometers in width along the plate boundaries, whilst deformation in continental lithosphere is more diffuse and can take place hundreds to thousands of kilometres away from plate boundaries (Isacks et al., 1968). Such internal deformation in certain regions of the continental lithosphere has been demonstrated more recently with geodetic techniques (Kreemer et al., 2003). The difference in behaviour of continental lithosphere has been ascribed to its greater buoyancy and lesser strength compared to oceanic lithosphere, resulting from the thicker crust in continental lithosphere (McKenzie, 1969; Molnar, 1988).

The classical example for diffuse intra-continental deformation is East Asia, where shortening, strike-slip and extensional deformation occur far from the India–Eurasia collision zone in the south and the subduction zones bordering East and Southeast Asia. The cause for this widespread deformation is generally thought to be the collision between India and Eurasia (Molnar and Tapponnier, 1975; Tapponnier et al., 1982; Davy and Cobbold, 1988; England and Houseman, 1988), but an important role for the western Pacific and Southeast Asian subduction zones has been implied more recently (Northrup et al., 1995; Fournier et al., 2004; Schellart and Lister, 2005; Royden et al., 2008).

## 5. Driving forces of plate tectonics and mantle convection

The theory of plate tectonics was initially developed as a kinematic theory describing the motions of plates with respect to one another. Very soon after development of this kinematic foundation, questions started to arise regarding the driving mechanism(s) of plate motion. Initially, it was thought that convection below the plates would provide the driving force. Convection in the sub-lithospheric mantle was hypothesized to occur separately from the motion of the plates above and the upper cool thermal boundary layer of the convective system was thought to be located in the upper part of the sub-lithospheric mantle (asthenosphere). Convective motion in the asthenosphere would exert a drag to the overlying lithosphere thereby driving plate motion. The mantle drag mechanism was already proposed by Holmes (1928) in order to explain the continental drift theory proposed by Wegener (1912, 1915), but has been discussed by numerous others including Ampferer (1906), Schwinner (1920), Vening Meinesz (1934), Griggs (1939), Runcorn (1962), Morgan (1972) and Turcotte and Oxburgh (1972).

In contrast to a driving force from below, Forsyth and Uyeda (1975), Solomon et al. (1975) and Chapple and Tullis (1977) proposed that the plates are driven by forces that are applied to the sides (plate boundaries). The main driving forces were thought to be the slab-pull force, where the slabs would pull the trailing subducting plates to which they are attached towards the trench, and the ridge push force, where the plates on either side of a spreading ridge are pushed away from these ridges. A third force, the trench suction force, was also proposed, which resulted from slab sinking and would drive overriding plates towards the subduction zone (Elsasser, 1971). The ridge push force acts perpendicular to the ridge axis and can be considered as the horizontal pressure gradient that arises from the cooling and thermal contraction of the oceanic lithosphere with increasing age, integrated over the area of the oceanic lithosphere (Lister, 1975; Meijer and Wortel, 1992). The magnitude of this force has been estimated at $2–3 \times 10^{12}$ N per meter ridge length (Harper, 1975; Lister, 1975; Parsons and Richter, 1980).

The slab-pull force results from the negative buoyancy of the subducting slab compared with the surrounding sub-lithospheric mantle. Slabs are negatively buoyant due to their higher average density compared to the ambient mantle (~80 kg/m$^3$ for 80 Ma oceanic lithosphere; Cloos, 1993). This high density is thought to have both a thermal component because the slabs are relatively cold, as well as a chemical component due to metamorphism of the basaltic oceanic crust into dense eclogite facies rocks (Cloos, 1993). The negative buoyancy force is thought to be transmitted to the surface because the slab and the lithosphere at the surface can act as a stress guide, where deviatoric tensional stresses can be transmitted from the slab across the hinge towards the surface part of the subducting lithosphere (Elsasser, 1967, 1971). The total negative buoyancy force of the slab is at least an order of magnitude larger than the ridge push force. For example, a 700 km long, 100 km thick 80 Ma slab (with density contrast $\Delta\rho = 80$ kg/m$^3$) has a negative buoyancy force of $5.5 \times 10^{13}$ N per meter trench length. But most of the negative buoyancy is thought to be absorbed by shear forces and slab-normal forces in the mantle resisting subduction and sinking of the slab (Forsyth and Uyeda, 1975; Davies and Richards, 1992; Schellart, 2004b) or in the subduction zone hinge (Conrad and Hager, 1999; Becker et al., 1999; Funiciello et al., 2003; Bellahsen et al., 2005). Also, it is not immediately clear how much of this force can actually be transmitted across a non-stationary subduction hinge.

Nevertheless, empirical data indicate that most plates that are attached for a large part of their circumference to subducted slabs have a relatively high velocity (Forsyth and Uyeda, 1975; Gripp and Gordon, 2002), underlining the importance of slab pull in driving tectonic plates. Furthermore, in most global reference frames (i.e. Pacific hotspot, Indo-Atlantic hotspot, global hotspot and no-net rotation), the major subducting plates (Pacific, Nazca, Cocos, Australia, Philippine, and Juan de Fuca) move trenchward (e.g. Chapple and Tullis, 1977; Chase, 1978; Gordon and Jurdy, 1986; Argus and Gordon, 1991; Gripp and Gordon, 2002; Schellart et al., 2008). In particular, in the Indo-Atlantic hotspot reference frame, even those plates with a small percentage of their circumference attached to a slab (e.g. Africa, South America, Arabia, and India) move trenchward with local velocity vectors on the subducting plate striking at a relatively high angle to the strike of the trench (Fig. 2) (Schellart et al., 2008).

## 6. Plate tectonics is mantle convection

More recently a common view emerged, in which plate tectonics and mantle convection are thought to be part of one and the same process, i.e. plate tectonics is the surface expression of mantle convection on Earth. Oceanic plates form the top cool thermal boundary layer of the convection cells (Davies and Richards, 1992; Davies, 1999; Tackley, 2000; Bercovici, 2003). The basis for this view was already outlined in the seminal paper from Elsasser (1971). In this view, plate motions and mantle convection are controlled primarily by the strong cold and dense thermal boundary layer at the top, i.e. the oceanic lithosphere, because the largest thermal gradient and the greatest source of potential energy reside in the oceanic lithosphere. The fundamental source of energy is the Earth's internal heat and most of this heat originates from the mantle (Davies, 1999; Leng and Zhong, 2009), but flow in the mantle is predominantly controlled and organized by the plates at the surface due to their greater strength. Some go further and argue in favour of a top-down control in which the sub-lithospheric mantle is essentially passive, whilst most potential energy in the oceanic plates dissipates within the plates themselves rather than in the mantle (Anderson, 2001; Hamilton, 2007). Regional-scale 3D models of progressive free subduction indeed imply that flow in the mantle is controlled primarily by the surface plates and their subducted slabs, but several studies also indicate that most viscous dissipation occurs in the ambient mantle, not in the plates and slabs themselves (Stegman et al., 2006; Capitanio et al., 2007, 2009). Note again that a large part of the negative buoyancy of subducted oceanic lithosphere has a chemical origin due to metamorphism of the basaltic crust into dense eclogite facies rocks at depths below ~40 km depth.

Separate from this plate tectonic style of mantle convection are the volcanic hotspot tracks, which are thought to be caused by thermal mantle plumes from deep within the mantle (Morgan, 1971). These mantle plumes are now thought to represent a secondary mode of mantle convection, originating from a lower hot thermal boundary layer due to bottom-up heating at the core–mantle boundary (Davies and Richards, 1992; Davies, 1999). Recent evidence from finite frequency tomography indicates that deep mantle plumes exist beneath a large number of known hot spots (Montelli et al., 2006); however, it should be noted that other tomography studies (e.g. Boschi et al., 2007) do not agree with these findings.

The actual driving mechanism for the plates is thus thought to be the negative buoyancy force of subducted slabs, which cause them to sink and actively pull the trailing plates into the mantle, at the same time also inducing flow in the mantle that drives both subducting plate and overriding plate towards the trench (Elsasser, 1971; Gurnis and Hager, 1988; Zhong and Gurnis, 1995; Conrad and Lithgow-Bertelloni, 2002; Schellart, 2004b; Sandiford et al., 2005; Capitanio et al., 2007). The potential energy stored in mid-ocean ridge systems is thus considered to be of minor significance. However, a major debate remains regarding the partitioning of the slab's potential energy in the different energy sinks. Two major energy sinks have been proposed: the subducting slab and the sub-lithospheric mantle. It has been argued that most of the energy is used to deform the slab at the subduction hinge, when the subducting plate is bent into the mantle (Houseman and Gubbins, 1997; Conrad and Hager, 1999; Becker et al., 1999; Funiciello et al., 2003; Bellahsen et al., 2005; Faccenna et al., 2007). Others propose that most of the energy is used to drive flow in the ambient mantle (Forsyth and Uyeda, 1975; Davies and Richards, 1992; Schellart, 2004b, 2009; Stegman et al., 2006; Capitanio et al., 2007; Nakakuki et al., 2008; Krien and Fleitout, 2008; Capitanio et al., 2009). Two of the crucial components in this debate are the magnitude of the bending radius and the effective viscosity ratio between the slab and ambient upper mantle.

The magnitude of viscous dissipation that occurs during bending of the slab scales with the inverse of the cube of the bending radius (Conrad and Hager, 1999; Buffett, 2006). Recent investigations indicate that the average bending radius on Earth is relatively large (~390 km) compared to what was previously assumed (~200 km) (Wu et al., 2008). This would reduce the average bending dissipation on Earth by a factor of ~1/8. Furthermore, laboratory and numerical studies on subduction zone geoids (Moresi and Gurnis, 1996), slab strain rates (Billen et al., 2003), subduction partitioning (Funiciello et al., 2008), slab geometries and trench geometries (Schellart, 2008b), global trench migration velocities (Schellart et al., 2008) and slab bending resistance (Wu et al., 2008; Stegman et al., 2010-this issue-b) suggest that in nature the effective viscosity ratio is relatively small ($1–7\times10^2$). In addition, it has been observed that with increasing relative strength of the slab, the bending radius increases as well, thereby maintaining the bending dissipation at a relatively low level (Capitanio et al., 2007; Schellart, 2008b, 2009; Capitanio et al., 2009). The relatively large bending radius and relatively low slab strength imply that viscous dissipation in the subducting slab is comparatively small with respect to viscous dissipation in the ambient mantle. This is indeed found in recent numerical and laboratory models, with dissipation in the slab not more than ~20% and dissipation in the ambient mantle of at least ~80% (Stegman et al., 2006; Capitanio et al., 2007; Krien and Fleitout, 2008; Schellart, 2009; Capitanio et al., 2009).

## 7. Mountain building and plate tectonics

Dewey and Bird (1970) were probably the first to explain mountain belts in a plate tectonic framework. They argued for the existence of two types of mountain belts: cordilleran mountain belts and collisional mountain belts. In their subdivision, cordilleran mountain belts form when a subduction zone develops near or at the continental passive margin to consume oceanic lithosphere from the oceanic side (e.g. Andes). Furthermore, such a mountain belt would form dominantly by thermal mechanisms related to the rise of a variety of magmas. In contrast, collisional mountain belts would form where a continental margin collides with another continent (e.g. Himalayas, Alps) or with an island arc (e.g. New Guinea, Taiwan), and such a mountain belt would form predominantly by mechanical processes (folding, thrusting, etc.).

### *7.1. Cordilleran (non-collisional) mountain belts*

Cordilleran mountain belts are also referred to as non-collisional mountain belts because they form at an ocean–continent subduction zone with continuous subduction of the oceanic lithosphere. The hypothesis of a dominantly thermal origin for Cordilleran orogens (Dewey and Bird, 1970) has been modified throughout the years, as it has become clear that overriding plate shortening and shortening-induced crustal thickening in the overriding plate can be significant. This is particularly so for the Andes mountains, for which shortening has been suggested to be the dominant, if not only, cause for crustal thickening and formation of mountain relief (e.g. Isacks, 1988). More recent work, however, indicates that crustal thickening by magmatic processes, and potentially detachment of the lower lithospheric root, might also be of importance for the formation of, in particular, the Central Andes topography (Sempere et al., 2008; DeCelles et al., 2009). Nevertheless, estimates for Cenozoic shortening in the Central Andes are significant and in the range of 250–350 km (Kley and Monaldi, 1998; Oncken et al., 2007) to 400 km (Arriagada et al., 2008), and this shortening has contributed significantly to crustal thickening. At other subduction zones around the globe, shortening and crustal thickening are much reduced or absent, except for potential frontal accretion at the subduction interface, or the overriding plate is characterized by backarc extension. Crustal thickening and formation of topographic relief in such continental or island arc settings results mostly from magmatic processes such as magmatic underplating, magmatic intrusion and volcanism.

The presence of significant shortening in the Central Andes, which contrasts with what is observed at most other subduction zones, has resulted in a proliferation of conceptual models that seek to explain this difference (see discussions in Oncken et al., 2007; Schellart et al., 2008). For example, it was suggested that subduction of relatively young oceanic lithosphere would induce overriding plate shortening (Molnar and Atwater, 1978), but the absence of shortening at other subduction zones that consume comparably young oceanic lithosphere (e.g. New Hebrides, Ryukyu) reduces the relevance of this model (Schellart et al., 2008). The presumed high friction due to subduction erosion along the subduction interface in the Central Andes has also been proposed (e.g. Lamb and Davis, 2003), but again, the absence of shortening at other subduction settings with comparable or even higher subduction erosion rates and presumably higher friction coefficients (e.g. Scotia, Tonga, and Ryukyu) makes the model unlikely.

Several statistical investigations point to a dominant role of overriding plate motion in determining overriding plate deformation (Jarrard, 1986; Heuret and Lallemand, 2005), although a more recent statistical investigation indicates that it only has a dominant role in a specific tectonic setting, namely in the centre of wide subduction zones (Schellart, 2008a). Nevertheless, two-dimensional numerical models of subduction indicate that significant trenchward overriding plate motion is required for compression and shortening to occur in the overriding plate (van Hunen et al., 2000; Buiter et al., 2001; Sobolev and Babeyko, 2005; Hampel and Pfiffner, 2006). Indeed, Silver et al. (1998) suggested that an increase in westward velocity of the South American plate at ~30 Ma coincided with inception of shortening in the Andes. However, more recent geological investigations indicate that shortening in the Central Andes began already at ~47 Ma (Oncken et al., 2007), whilst phases of earlier shortening along the Andes have also been documented (Sempere et al., 2008). Also, the 2D numerical models do not explain the trench-parallel variation in shortening along the Andes, which is maximum in the centre and decreases progressively towards the north and south (Kley and Monaldi, 1998; Oncken et al., 2007; Arriagada et al., 2008). This implies that the Central Andes has experienced, and continues to experience, larger resistance to westward motion compared to regions to the north and south.

A potential explanation for trench-parallel variation in shortening in the Andes could be offered by incorporating the third, trench-parallel, dimension, which was ignored in the 2D numerical studies mentioned above. Seismic anisotropy studies led Russo and Silver (1994, 1996) to suggest that in the Bolivian region a mantle stagnation point developed on the ocean side of the Nazca slab. This stagnation point resists lateral slab migration, and as such the overriding plate is effectively colliding with a relatively immobile mantle wall in this region. Schellart et al. (2007) provided support for this conceptual model with 3D numerical models of progressive free subduction. In these models subduction of a wide slab caused formation of a mantle stagnation zone in the centre behind the slab, which resists trench migration and would thereby be able to support a large mountain belt like the Central Andes. As such, westward motion of the South American plate would be accommodated by shortening in the centre and by progressively less shortening but progressively more trench retreat in the north and south. A global statistical analysis of overriding plate shortening at subduction zones lends additional support to this conceptual model (Schellart, 2008a). Although this model provides a better explanation for the spatial distribution of shortening along the Andes, it currently provides no explanation for the onset of mountain building in the Andes in the Early Cenozoic. Indeed, continuous subduction has been active along the western boundary of the South American plate since at least the Early Mesozoic (Oncken et al., 2007; Sempere et al., 2008).

Another problem regards the driving force(s) of Andean orogeny, which basically comes down to the forces that drive the South American plate westward. Three major contributors can be thought of (Meijer and Wortel, 1992; Coblentz and Richardson, 1996), namely: (1) ridge push from the mid Atlantic spreading ridge; (2) slab pull from the Scotia slab and the Lesser Antilles slab; and (3) tractions at the base of the South American plate most likely induced by the sinking Nazca slab. It remains to be demonstrated with fully dynamic models if these forces provide sufficient drive for westward motion of the South American plate and formation of the Andes mountain belt.

### 7.2. Collisional mountain belts

Collisional mountain belts form where two continents collide or where a continent collides with an island arc. Typical examples of continent–continent collisions include the Alps, which resulted from collision of the Adriatic promontory of Africa with southern Europe, and the Himalayas, which resulted from collision of the Indian continent with southern Asia. Typical arc–continent collisions include the Taiwan mountain belt, which resulted from collision of the Luzon arc with eastern Asia, and the New Guinea mountains, which originated from collision of the Finistre arc with the northernmost part of the Australian continent.

A large component of research into mountain belts has focussed on estimating the amount and timing of shortening in the direction perpendicular to the axis of the orogen and on the cross-sectional structure of mountain belts (e.g. Price, 1981; ECORS Pyrenees team, 1988; Pfiffner et al. 1990; Schmid et al., 1996; Teixell, 1998; Fitzgerald et al., 1999; Friberg et al., 2002). Such findings might then be put in a plate-tectonic framework, where estimates of amount and timing of shortening can be related and compared to estimates for convergence in plate tectonic reconstructions. For example, the timing and variation of shortening as documented along the axis of the Pyrenees mountain belt (e.g. Labaume et al., 1985; ECORS Pyrenees team, 1988; Muñoz, 1992; Teixell, 1998; Fitzgerald et al., 1999) are comparable to estimates from plate tectonic reconstructions (e.g. Rosenbaum et al., 2002a).

Another important branch of research into mountain belts has focussed on their three-dimensional nature. A problem of particular interest regards their plan-view curvature, and Carey (1955) was one of the first to try and explain this from a mobilistic perspective. Current research investigates whether segments of mountain belts have experienced vertical axis rotations in order to establish if mountain belt curvature is of primary origin (non-rotational arcs), rotational origin (i.e. oroclines, orogens that were originally linear but subsequently curved) or progressive origin (curvature is progressively acquired during formation of the mountain belt) (e.g. Marshak, 1988; Costa and Speranza, 2003; Weil and Sussman, 2004; Soto et al., 2006).

Other 3D complications for collision zones involve the potential orogen-parallel transport of crustal and/or lithospheric material, such as during lower crustal flow (e.g. Bird, 1991; Royden et al., 1997, 2008) or during extrusion/escape tectonics (e.g. Tapponnier et al., 1982; Davy and Cobbold, 1988; Ratschbacher et al., 1991). During extrusion or escape tectonics, large lithospheric blocks or microplates are squeezed out of the collision zone along lithospheric-scale strike-slip faults towards regions of reduced confinement. This mechanism was first proposed by McKenzie (1972) for the Bitlis–Anatolia–Aegean–Hellenic system and has later been applied to the India–Eurasia collision zone (Molnar and Tapponnier, 1975) and Alpine–Carphatian system (Ratschbacher et al., 1991). Many new insights have been obtained from scaled laboratory models (e.g. Tapponnier et al., 1982; Davy and Cobbold, 1988; Ratschbacher et al., 1991; Martinod et al., 2000; Sokoutis et al., 2000; Fournier et al., 2004; Cruden et al., 2006; Schueller and Davy, 2008) and numerical thin viscous sheet models (e.g. England and Houseman, 1988; Houseman and England, 1993) simulating collisional tectonics and extrusion tectonics. These include the importance of a free boundary that allows escape of lithospheric-scale blocks, buoyancy forces, the interplay between thrust, strike-slip and normal faults and the factors controlling localized versus distributed deformation.

A fundamental question regarding the origin of collisional mountain belts relates to the forces that drive their formation. This is not a trivial question, considering that such mountain belts are characterized by considerable topography, which stores massive amounts of gravitational potential energy. As such, the driving forces not only have to overcome the mechanical resistance of the lithosphere to break and deform in the collisional zone, they also have to work against gravity (Molnar and Lyon-Caen, 1988). Thus, the buoyancy forces stored in a mountain belt counteract the driving forces of mountain building such as ridge push. For example, the potential energy stored in a lithospheric column of thickened crust of the Tibetan plateau is much larger than that of a spreading ridge in the Indian ocean, so it is not immediately clear how ridge push could drive continued mountain building in the Himalayas–Tibet collision zone. One solution might be that during shortening and crustal thickening, the buoyant continental crust is underlain by a thickened and dense lithospheric mantle root, and as such, the gravitational potential energy integrated from the surface down to the compensation depth at the base of the lithosphere is much reduced (Molnar et al., 1993). In such a scenario, work against gravity is much reduced as well. Formation of a significant mountain belt with a high plateau like Tibet will only form once the dense lithospheric root is convectively removed, after which the thickened crust pops up and might start to collapse (England and Houseman, 1989; Molnar et al., 1993).

Another solution for mountain building like the Himalayas–Tibet might be that the total ridge length of the Indo-Australian composite plate is several times larger than its total collisional length (Sandiford et al., 1995). As such, the total ridge push force integrated over the entire ridge length might be comparable to the total collisional resistance along the Himalayas collisional boundary and the resistance in the New Guinea region. Furthermore, slab-pull forces from the slabs subducting along the Andaman–Sumatra–Java and New Britain–San Cristobal–New Hebrides subduction zones provide an additional force for north to northeast-directed motion of the Indo-Australian composite plate (e.g. Sandiford et al., 2005). Finally, slab segments have been imaged with seismic tomography and are delineated by earthquake hypocentres below the collision zones of the Himalayas (Van der Voo et al., 1999; Replumaz et al., 2004, 2010-this issue), Burma (Bijwaard et al., 1998) and the Banda arc (Richards et al., 2007; Ely and Sandiford, 2010-this issue), which provide an additional north to northeast-directed pull on the Indo-Australian composite plate.

The forces required for construction of smaller mountain belts such as the (European) Alps and the New Zealand Alps could be explained in a similar fashion as described in the previous paragraph. For example, the collisional plate boundary length of the New Zealand Alps is more than an order of magnitude smaller than the spreading ridge length of the Pacific plate and more than an order of magnitude smaller than the cumulative subduction zone length of the Pacific plate. Thus, the forces that arise from the East Pacific Rise and the western Pacific subduction zones drive the Pacific plate westward and are sufficient to cause mountain building in New Zealand. A similar case can be argued for the Alps, with a short collisional plate boundary length compared to the total circumference of the African plate.

A complicating factor for both collisional and non-collisional mountain belts is that they are mostly located above sea level and as such interact with the atmosphere and hydrosphere and are subject to weathering and erosion. On the one hand it has been suggested that the formation of mountain belts such as the Himalayas affects the regional climate and potentially the global climate (e.g. Molnar et al., 1993). On the other hand it has been argued that climate and erosion affect the kinematics and dynamics of orogenesis (e.g. Beaumont et al., 1996; Chemenda et al., 2000; Beaumont et al., 2001; Cruz et al., 2008) such as for the New Zealand Alps and the Himalayas through focussed erosion on one side of the mountain belt.

Another topic of interest for both collisional and non-collisional mountain belts regards the potential episodicity of mountain building, such as alternating phases of shortening (construction) and extension (destruction) (e.g. Lister and Forster, 2009), or episodic phases of magmatic growth of the continental arc (e.g. DeCelles et al., 2009). Whilst it is generally agreed that plate motions are relatively smooth and continuous, this does not imply mountain building at the interface between converging plates is also smooth and continuous (Lister et al., 2001). Indeed, structural geological, metamorphic and geochronological data from several orogens imply that mountain building is episodic and might be characterized by alternating phases of shortening and extension (Lister et al., 2001; Rawling and Lister, 2002; Lister and Forster, 2009).

Related to the question of episodic behaviour of mountain belts is the mechanism that is responsible for the destruction of mountain belts and exhumation of high-pressure rocks from great depths. Proposed mechanisms include gravitational collapse (e.g. Dewey, 1988; Platt and Vissers, 1989), extrusion of a crustal wedge in a subduction channel (Chemenda et al., 1996, 2000; Beaumont et al., 2001), or the episodic rollback of the subduction zone hinge causing extension of the overriding lithosphere and exhumation along extensional detachments (Lister et al., 2001; Lister and Forster, 2009). Constraints on such conceptual models will come from detailed structural and metamorphic investigations of the high-pressure rocks and the deformed zones in which they occur, and estimates of exhumation rates. For example, some recent estimates of exhumation rates are very high, with ~3.4 cm/yr for ultrahigh-pressure rocks in the Alps (Rubatto and Hermann, 2001) and ~1.7 cm/yr for high-pressure rocks in eastern Papua New Guinea (Baldwin et al., 2004). Geodynamic models need to investigate the physical plausibility of different conceptual models and test whether model exhumation rates of high-pressure rocks and the structural setting in which they occur are comparable to those documented in nature.

## 8. Plate boundary kinematics

After the basic principles of plate tectonic theory were developed in the 1960s, it was soon realised that not only do the plates move across the globe, the plate boundaries (plate margins), and triple junctions for that matter, move across the globe as well. One of the first to realise this was Elsasser (1971), who noted that the Atlantic Ocean, which stretches from the North Pole to close to the South Pole, is opening up in an east–west fashion. This implies that on a constant-radius Earth the Pacific Ocean must be shrinking. This again would indicate that the subduction zones in the Eastern Pacific and those in the Western Pacific are moving towards each other, and thus, that subduction zone plate boundaries migrate with time. Apart from subduction zones, it was also realized that spreading ridges and transform faults migrate laterally with time (Kaula, 1975).

As spreading ridges and transform faults are geometrically symmetrical in a plane perpendicular to their strike, and because the upwelling beneath spreading ridges is essentially passive (Davies, 1999), one can expect them to migrate in either direction. Subduction zones, however, are geometrically asymmetric in a plane perpendicular to their strike, and the downwelling at a subduction zone is not passive: the sinking slabs are thought to be driven by their own negative buoyancy with respect to the ambient mantle (see Section 5). Considering that the negative buoyancy force of the slab is directed vertically downward and that slabs naturally dip towards the mantle wedge, it was initially thought that subduction zone migration occurs only in a backward fashion, i.e. towards the subducting plate side (trench/hinge/slab retreat/rollback or retrograde slab motion) (Elsasser, 1971; Havemann, 1972; Molnar and Atwater, 1978; Garfunkel et al., 1986; Hamilton, 1988). The effects of such trench migration were investigated in laboratory models and numerical models in which trench motion was kinematically prescribed (e.g. Griffiths et al., 1995; Guillou-Frottier et al., 1995; Christensen, 1996). These models showed that rapid trench retreat causes slabs to be horizontally draped on top of the upper–lower mantle transition zone at ~670 km depth, whilst a fixed trench causes slabs to sink more readily through the transition zone into the lower mantle.

In the late 1970s and 1980s, calculations of plate boundary kinematics indicated that for a minority of subduction zones trench migration occurs in a forward motion (i.e. trench/hinge/slab advance) (Chase, 1978; Carlson and Melia, 1984; Jarrard, 1986). Notwithstanding such kinematic support for both backward and forward subduction zone migration, early dynamic models of progressive free subduction showed that slabs preferentially retreat (Jacoby, 1973, 1976; Kincaid and Olson, 1987). Slab advance could only be achieved by producing a strong active upwelling at the trailing edge of a subducting plate, thereby pushing it forward (Jacoby, 1976). Predominance of slab retreat was also observed in more recent models with a subducting plate fixed at its trailing edge (Funiciello et al., 2003; Schellart, 2004a; Enns et al., 2005) and models with a free trailing edge (Schellart, 2004a; Enns et al., 2005).

More recently, Heuret and Lallemand (2005) suggested that subduction zone trench migration is approximately equally partitioned between trench retreat (52%) and trench advance (48%) for most active subduction zones on Earth using a Pacific hotspot reference frame from Gripp and Gordon (2002). The equal partitioning would imply that there should be as many backward draping slab geometries (with a sub-horizontal slab segment at the transition zone below the overriding plate) as forward draping (roll-over) slab geometries (with a sub-horizontal slab segment at the transition zone below the subducting plate). This is contrary to observations for subduction zones, where several backward draping slab geometries exist (e.g. Tonga, Japan, Izu–Bonin, and Calabria), but no forward draping slab geometries exist. Furthermore, it later became clear that the equal partitioning breaks down when using a complete global dataset of subduction zones and updated rotation parameters for plates and plate boundary deformation (62% retreat and 38% advance) (Schellart et al., 2008). The equal partitioning breaks down even further (75% retreat and only 25% advance; Schellart et al., 2008) when using a more recent, and geodynamically more realistic global reference frame (Indo-Atlantic hotspot). In this reference frame all plates that are attached to subduction zones move towards these subduction zones (Fig. 2), and the global net westward rotation is much reduced compared to the Pacific hotspot reference frame. Such a reduced net westward rotation is in accordance with recent predictions from global convection models (Becker, 2008).

Recent 3D dynamic models of progressive subduction have shown that slabs can both retreat and advance (Funiciello et al., 2004; Bellahsen et al., 2005; Schellart, 2005; Schellart et al., 2007; di Giuseppe et al., 2008; Schellart, 2008b). Such models found that important controls on the trench kinematics are the boundary condition at the trailing edge of the subducting plate (Funiciello et al., 2004; Schellart, 2004a, 2005), the width (trench-parallel extent) of the slab (Schellart, 2004a; Stegman et al., 2006; Morra et al., 2006; Schellart et al., 2007) and the relative strength of the subducting slab with respect to the ambient upper mantle (Capitanio et al., 2007; Schellart, 2008b; Funiciello et al., 2008). These findings are corroborated by more recent 3D numerical models of progressive subduction that illustrate the importance of boundary conditions and slab width on trench migration velocities (Stegman et al., 2010-this issue-a) as well as the importance of the slab to upper mantle viscosity ratio and slab–upper mantle density contrast (Stegman et al., 2010-this issue-b). In particular, these works indicate that for fully dynamic models reflecting Earth-like conditions with respect to viscosity ratios and slab widths, trenches preferentially retreat, except in the centre of wide subduction zones (Stegman et al., 2010-this issue-a,b) (see also Schellart et al., 2007; Schellart, 2008b; Ozbench et al., 2008) or when a buoyant feature like an oceanic plateau enters the trench (Mason et al., 2010-this issue). As mentioned before, predominance of trench retreat is indeed expected from slab geometries for the active subduction zones on Earth, which show a lack of slab roll-over geometries that would form if trench advance was more dominant.

## 9. Slab sinking kinematics

The observation that subduction zone trenches and their hinges migrate with time has important consequences for the sinking kinematics of slabs. Trench migration, as well as slab steepening or flattening, implies that slabs sink at an angle with respect to their dip plane. During the advent of plate tectonics in the 1960s and the following decade or two it was apparently assumed by many that slabs sink into the mantle following a path parallel to the dip of the slab. Indeed, many textbooks show subduction zones with slabs sinking in a down-dip direction. Also, mantle wedge corner flow models generally assumed only down-dip slab motion (e.g. McKenzie, 1969; Bodri and Bodri, 1978; Toksöz and Hsui, 1978), with the models from Garfunkel et al. (1986) being an important exception. A third example of the assumption of down-dip slab motion is the well-known global kinematic flow models from Hager and O'Connell (1978).

In the Hager and O'Connell (1978) models, flow is driven by kinematic boundary conditions applied to the surface plates, whilst buoyancy forces were ignored. A very good agreement was found between the alignment of slabs as determined from Wadati–Benioff zones and upper mantle flow directions in the vicinity of those slabs. As discussed by King (2001) this leaves a rather odd paradox because it is clear that slabs are denser than their surrounding (Hager, 1984; Cloos, 1993) and provide a major driving force for plate motions (Elsasser, 1971; Forsyth and Uyeda, 1975; Chapple and Tullis, 1977). But inclusion of buoyancy forces for slabs would increase the vertical component of slab motion relative to the purely kinematic flow, and thus would decrease the alignment between slab dip and mantle flow direction. This paradox has lead some (e.g. Billen, 2008) to suggest that slabs must be very strong in order to counter the force of gravity thereby reducing the vertical component of sinking.

A solution to this apparent paradox appears to be hidden in the inherent assumption made by Hager and O'Connell (1978) in their comparison between observation and models that slabs sink down-dip, parallel to their own plane. This very basic assumption is incorrect. Elsasser (1971) already proposed in the early 1970s that on Earth slabs sink at an angle with respect to their dip angle, and several others soon followed suit making similar claims (e.g. Havemann, 1972; Molnar and Atwater, 1978). The fact that the force of gravity is oriented vertically downward would suggest that slabs sink subvertically, so at an angle that is steeper than their dip (Hamilton, 1988). This hypothesis is supported by fluid dynamic considerations for the sinking of planar objects into a fluid at low Reynolds number (Dvorkin et al., 1993). The concept that the sinking trajectory of a slab is at an angle with respect to the plane of the slab is further supported by laboratory models of subduction (e.g. Jacoby, 1976; Schellart, 2004a,b, 2005, 2008b). These models point towards even greater complexity, because the slab is not just a planar body sinking in isolation. Slabs are connected to the trailing subducting plate at the surface, and the velocity at which this surface plate moves towards the trench will modify the sinking behaviour of the slab (Schellart, 2005). The laboratory models show that during trench retreat slabs sink at a high angle with respect to the dip of the slab (30–50°), and the sinking direction can be backward (oceanward) to subvertical. During trench stability slabs sink subvertically, whilst during trench advance slabs sink forward (towards the mantle wedge) at an angle that is smaller than the dip of the slab.

## 10. Four-dimensional aspects of convergent plate boundaries

Recent developments in structural geology and tectonics, geophysics and geodynamic modelling indicate that convergent margins are complex settings with an intrinsically three-dimensional geometry that evolves with time. Indeed, subduction zones and collision zones have a limited trench and orogen-parallel extent, and display various degrees of contortion and curvature in plan-view (Fig. 2). The first plate tectonic hypothesis for subduction zone curvature was offered by Frank (1968), which implied an incipient, static origin. Later hypotheses explained such curvature with subduction of buoyant features such as aseismic ridges and plateaus (e.g. Vogt, 1973; Hsui and Youngquist, 1985; Martinod et al., 2005; Wallace et al., 2009; Mason et al., 2010-this issue), with toroidal return flow around lateral slab edges (e.g. Jacoby, 1973; Schellart, 2004a; Morra et al., 2006; Schellart et al., 2007; Loiselet et al., 2009), or a combination thereof. In addition to trench and arc curvature, subducted slabs show variations in slab dip angle and bending curvature at the subduction hinge and at depth (Jarrard, 1986; Yamaoka et al., 1986; Gudmundsson and Sambridge, 1998; Wu et al., 2008), folding of slabs at and below the transition zone (Griffiths et al., 1995; Guillou-Frottier et al., 1995; Schellart, 2005; Ribe et al., 2007), as well as all sorts of discontinuities including subvertical slab edges, seismic gaps, slab kinks and slab tears (e.g. Millen and Hamburger, 1998; Wortel and Spakman, 2000; Govers and Wortel, 2005; Lin et al., 2007; Rosenbaum et al., 2008; Gasparon et al., 2009; Yang et al., 2009; Ely and Sandiford, 2010-this issue). This all points towards great four-dimensional complexity of subduction zones.

Recent advances in both earthquake location and seismic tomography have further helped to refine our understanding of subduction zones. Highly accurate relative location techniques such as the double difference scheme of Waldhauser and Ellsworth (2000), which can provide more than an order of magnitude improvement in estimated hypocenter uncertainties compared to catalogue locations, have allowed fine scale slab structure to be much more accurately delineated by the pattern of seismicity (e.g. Rietbrock and Waldhauser, 2004). Coupling these location techniques with local earthquake tomography has also allowed a commensurate improvement in the detail of recovered velocity structure (Zhang et al., 2004). The recent advent of finite frequency tomography, which can account for scattering effects in the Earth that are ignored by conventional geometric ray-based tomography, promises to deliver sharper images of subducting slabs, as it already has done (though somewhat controversially) with mantle plumes (Montelli et al., 2004).

In addition to technique development, improvements in seismic imaging can also be attributed to the increasing availability of high quality digital seismic datasets, and more powerful computers. For instance, recent high resolution tomographic imaging beneath the Japan islands (Matsubara et al., 2008) uses over five and half million traveltimes from 77,730 earthquakes and constrains a 3-D model comprising over half a million parameters. The resulting velocity images and relocated hypocentres provide an unprecedented view of the detailed slab geometry beneath this region. In global tomography, the inversion of huge datasets and the use of adaptive parameterizations, which allow more detail in highly seismogenic zones, has permitted high resolution images of subducted slabs to be produced throughout the full mantle thickness (e.g. Burdick et al., 2008). Fig. 4 shows a vertical slice through a recent P-wave traveltime tomography model in the region of the Cascadia subduction zone generated from global network and USArray data.

The complex 3D slab geometries that are currently imaged with seismic techniques, such as slabs that flatten out at the transition zone, folded slab piles and steep slabs that penetrate the transition zone, are reproduced to a large extent in fully dynamic 3D models of progressive free subduction (Jacoby, 1976; Kincaid and Olson, 1987; Schellart, 2004a, 2005; Funiciello et al., 2004; Stegman et al., 2006; Schellart et al., 2007; Capitanio et al., 2007; di Giuseppe et al., 2008; Schellart et al., 2008; Schellart, 2008b; Stegman et al., 2010-this issue-a,b; Hale et al., 2010-this issue; Mason et al., 2010-this issue).

Slabs are surrounded by heterogeneous mantle that points to complex three-dimensional flow patterns in the mantle as implied by seismic studies (Russo and Silver, 1994; Peyton et al., 2001; Müller et al., 2008; Zandt and Humphreys, 2008). Two standard seismic techniques that provide some insight into mantle flow are shear wave splitting and anisotropy tomography. In the former case, shear waves can split into two orthogonally polarized components on encountering an anisotropic medium; the differential arrival time between the waveforms provides insight into the strength and orientation of the anisotropy. The underlying cause of upper mantle anisotropy is the lattice preferred orientation of olivine which is strain-induced. Due to the path integral nature of shear wave splitting, it is difficult to unravel complex 3-D flow patterns from splitting results alone, although much has been learnt from this relatively simple technique (e.g. Long and Silver, 2008). In the last few years shear wave splitting tomography (e. g. Zhang et al., 2007) has started to become popular as a means of trying to spatially distribute anisotropy inferred from shear wave splits. More conventional anisotropy tomography tends to exploit an abundance of data coverage in an attempt to resolve both isotropic and anisotropic parameters. The under-determined nature of seismic tomography problems has generally resulted in relatively simple forms of anisotropy being assumed. For example, one of the first studies to reveal upwelling and downwelling features associated with slab subduction made the assumption of radial anisotropy (Nataf et al., 1984). The assumption of azimuthal anisotropy (e.g. Tanimoto and Anderson, 1984) is more attuned to contemporary plate motion, although "vectorial tomography" (Montagner and Tanimoto, 1991), which is able to resolve anisotropy associated with both vertical and lateral flow, is best suited to the examination of mantle dynamics associated with subduction zones. Complex 3D mantle flow patterns as implied by seismic anisotropy studies are reproduced to various degrees with geodynamic modelling, including toroidal-type flow

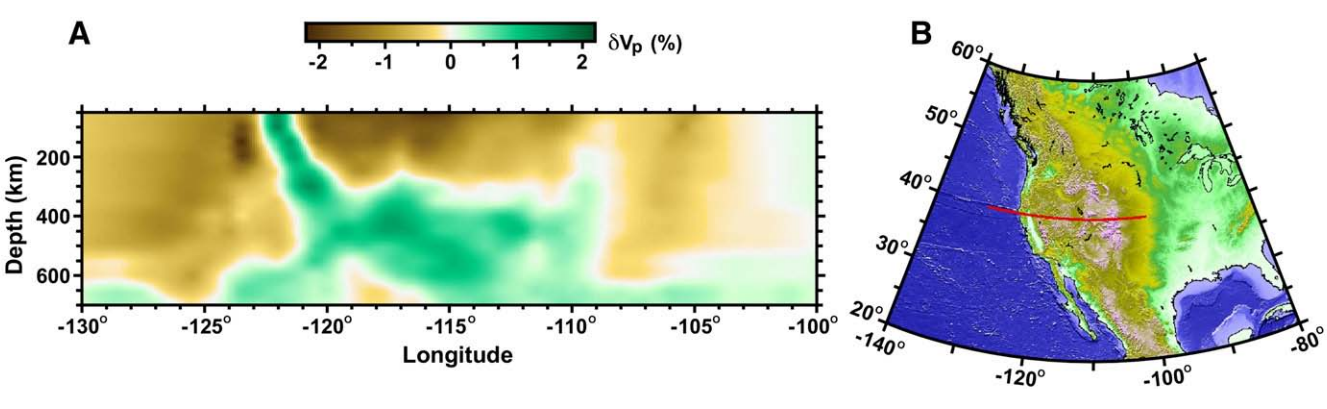


**Fig. 4.** Tomographic cross-section showing seismic velocity image of the Cascadia subduction zone at 40° north with eastward subduction of the Juan de Fuca plate, along with a location map (red line shows extent of cross-section). Velocities are plotted as percentage perturbations from the global reference model ak135 (Kennett et al., 1995). The image is based on a P-wave tomographic model constructed by Scott Burdick using global network and USArray data (Burdick et al., 2008). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

around lateral slab edges due to lateral slab migration (Buttles and Olson, 1998; Kincaid and Griffiths, 2003; Schellart, 2004a; Funiciello et al., 2006; Stegman et al., 2006; Schellart, 2008b; Honda, 2009) and trench-parallel flow in the mantle wedge due to curved slab geometry (Kneller and van Keken, 2007).

Tectonic reconstructions show that the shapes of collision zones and subduction zones have evolved over millions of years and continue to evolve with time (e.g. Carey, 1955; Malinverno and Ryan, 1986; Hall, 1996; Lonergan and White, 1997; Rosenbaum et al., 2002b; Hall, 2002; Replumaz et al., 2004; Miller et al., 2006; Martin, 2006, 2007; Richards et al., 2007; Arriagada et al., 2008; Schellart et al., 2009; Replumaz et al., 2010-this issue), thereby pointing towards the 4D nature of convergent plate margins. Such tectonic reconstructions indicate that subduction zones might form at some stage in the geological past and evolve into mature subduction zones, with some eventually terminating by turning into an obduction zone, arc–continent collision zone, or a continent–continent collision zone. Such a final stage of subduction might culminate in the detachment of the slab from the surface plate, a process that has been reproduced in laboratory models (Chemenda et al., 2000; Regard et al., 2003; Boutelier et al., 2004; Regard et al., 2005) and numerical models of subduction (Gerya et al., 2004), and has been suggested for subduction zones and collision zones in the Mediterranean region (Wortel and Spakman, 2000; Lei and Zhao, 2007; Zor, 2008), the Alpine–Himalayan chain (Van der Voo et al., 1999; Hafkenscheid et al., 2006; Replumaz et al., 2010-this issue), the Southwest Pacific region (Hall an Spakman, 2002; Schellart et al., 2009) and Central America (Ferrari, 2004).

## 11. Conclusions

In this review paper we have presented a historical outline of the development of geological theories for large-scale deformation of the Earth, and we have discussed more recent developments regarding the understanding of the structure of, and geodynamic processes operating at, convergent plate margins. From this review it is clear that structural geology, geophysics and geodynamic modelling have been critical in continuously advancing this understanding. Structural geological investigations of mountain belts dating back to the 18th century provided the first irrefutable evidence of large-scale deformation in mountain belts, whilst geodynamic modelling dating back to the 19th century demonstrated that such deformation implies large-scale horizontal motions, and thus, that the Earth's crust is mobile on geological time scales. Geophysical and geological investigations of the ocean floor dating from the mid 20th century have provided the basis on which the plate tectonic theory was developed. Seismological investigations starting in the early 20th century presented firm evidence of underthrusting and sinking of the oceanic lithosphere below overriding plates into the upper mantle at deep sea trenches, which mark subduction zone plate boundaries. Such research later established that in some cases slabs sink down into the lower mantle.

Structural geological investigations of convergent plate margins now focus on the potential episodicity of mountain building and destruction, the mechanisms responsible for such episodic behaviour and the mechanisms and rates of exhumation of high-pressure metamorphic rocks. Regional-scale geodynamic modelling of collisional margin dynamics can demonstrate the physical plausibility of various conceptual models of mountain building, destruction and exhumation, such as indentor tectonics, gravitational collapse, extrusion tectonics, slab rollback, convective delamination and slab detachment. They further provide quantitative insight into the influence of the different physical parameters on orogenic systems.

Regional and global geodynamic investigations of subduction systems now focus on questions such as the strength of the subducted slab, the role of the overriding plate, the major energy sinks in the system (e.g. ambient mantle, slab, subduction zone hinge, plate boundary interface), and the partitioning of the subduction velocity into the two components of trenchward subducting plate motion and trench retreat. The problem of subduction partitioning has an additional complicating factor in that it depends on the choice of reference frame in which these two velocity components are calculated, and as such, raises the question as to which, if any, reference frame on Earth (e.g. Pacific hotspot, Indo-Atlantic hotspot, no-net-rotation, minimal viscous dissipation) is most suitable to function as an absolute reference frame.

Global mantle convection models have progressively become more sophisticated and recent models produce planform mantle convection patterns that are increasingly realistic with respect to the plate-tectonic-style of mantle convection on Earth with rigid plate-like features at the top and linear downwellings similar to subduction zones. Major challenges include the one-sidedness of subduction zones, which is still poorly replicated in global convection models. Nevertheless, with continuous progress it becomes possible to answer questions as to why mantle convection on Earth is presently in a plate tectonic (mobile lid) regime, whilst other rocky planets such as Mars and Venus are evidently not.

Development of new techniques in seismology and the increasing volumes of high quality seismic data that are becoming readily available has resulted in more accurate determination of hypocenter locations, thereby providing more accurate delineation of Wadati–Benioff zones, more detailed imaging of slabs in mantle tomography models and more accurate mapping of seismic anisotropy in the mantle.

Major advances into the understanding of the three-dimensional interaction between the overriding plate, the subducting/underthrusting plate and the underlying mantle at convergent plate margins will likely result from multidisciplinary investigations that include, but are not limited to, structural geology, solid-Earth geophysics and geodynamic modelling.

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