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Yin and yang of continental crust creation and destruction by plate tectonic processes

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Earth's continental crust today is both created and destroyed by plate tectonic processes, a balance that is encapsulated by the traditional Chinese concept of *yin–yang*, whereby dualities act in concert as well as in opposition. *Yin–yang* conceptualizations of crustal growth and destruction are mostly related to plate tectonics; both occur mostly at subduction zones, by arc magmatic creation and by subduction removal. Crust is also created and destroyed by processes unrelated to plate tectonics, including losses by lower crust foundering and additions at hotspots. At present, creation and destruction of continental crust is either in balance (~3.2 km^3/year, or 3.2 AU) or more crust is being destroyed than created; the uncertainty comes from unknown deep losses of continental crust at collision zones and due to lower crustal foundering. The *yin–yang* creation–destruction balance changes over a supercontinent cycle, with crustal growth being greatest during supercontinent break-up due to high magmatic flux at new arcs and crustal destruction being greatest during supercontinent amalgamation due to subduction of continental material and increased sediment flux due to orogenic uplift. These conclusions challenge the widely held view that continental crust volume has increased over time due to plate tectonic activity; it is just as likely that this volume has decreased.

Keywords: continental crust; plate tectonics; supercontinent cycle; subduction

Introduction

Continental crust covers approximately one-third of Earth's surface, mostly (75%) lying above sea level. Its existence and stability allowed life on land to evolve from that in the sea, making possible among other things humans and civilization. Continental crust as old as 4 Ga is known (Bowring and Williams 1999) and great tracts of ultrastable ancient crust or cratons comprise the nuclei of all continents today. Nd isotopic studies indicate that the mean age of the continental crust is about 2.0 Ga (Hawkesworth and Kemp 2006), more than an order of magnitude greater than the age of the oldest ocean crust (Jurassic, ~180 Ma; but note that Müller *et al.* 2008 argue that oceanic crust as old as 230–270 Ma may exist in the Ionian and eastern Mediterranean basins). In addition to its much greater age, the greater thickness (~40 km) and lower density (~2.75 g/cc) of continental crust contrasts with thinner (~7 km), denser (~2.90 g/cc) oceanic crust. These differences contribute to the relative stability of continental crust and ephemeral nature of oceanic crust.

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Mature continental crust is compositionally layered and can be subdivided into upper continental crust (~20 km thick) of approximately granodioritic composition and lower, mafic crust (~20 km thick), separated by a Conrad discontinuity that is well to weakly formed (Christensen and Mooney 1995; Rudnick and Fountain 1995).

In contrast to the relative permanence of continental crust, the continents themselves reconfigure constantly, as different crustal tracts rift apart, collide, move parallel to the margin in orogenic streams (Redfield *et al.* 2007; also see the animated tectonic reconstructions by Tanya Atwater at http://emvc.geol.ucsb.edu/animations/quicktime/sm02PacNoAmflat.mov), and recoalesce as parts of a grand supercontinent cycle (Condie 2002). This accompanies the opening and closure of the much more transitory ocean basins. Many aspects of continental crust evolution are controversial, including whether the lower crust is recycled to the mantle differently than the upper crust and over how much of Earth history plate tectonics has operated and so could be the dominant mode of continental crust formation (e.g. Condie and Kröner 2008; Stern 2008; Ernst 2009).

Today there are approximately 7 billion cubic kilometres of continental crust (Cogley 1984), but in the past there may have been more crust or less. Most Earth scientists assume that continental crust volume has been growing over Earth history (see review and critique by Armstrong 1991), but when the question 'has continental crust volume increased or decreased with time?' is addressed thoughtfully, it must be admitted that there is much that is not known. It is clear from truncations of ancient orogenic belts (Dickinson 2009) and the presence of >4.0 Ga zircons that much Precambrian continental crust has been destroyed (Harrison 2009). It seems likely that the early Earth, like the Moon and other silicate planets, had an early sialic or feldspathic crust that covered the entire planet and that the present continental crust may be largely recycled from this primordial crust, even though remnants of it are hardly preserved (Bowring and Housh 1995). Such questions and controversies promise fruitful avenues of future solid Earth science research in their own right, even as they provide important perspectives on Earth's tectonic history and how plate tectonic processes build and destroy continental crust today.

This article considers one piece of the puzzle by critically examining how continental crust is produced and destroyed today. By 'today', we mean the recent geologic past, essentially since the oldest preserved modern oceanic crust formed (~180 Ma), a time period that is usefully approximated by the Cenozoic and Mesozoic eras. This allows us to disentangle how continental crust is created and destroyed today from the related important questions of how crust was created and destroyed in the past, especially during Precambrian time, whether or not plate tectonics operated then, and what was Earth's preplate tectonic style? Even though we focus on the modern Earth, our conclusions should be useful for understanding ancient continental crust production and destruction. Times when rates of ancient crust formation and destruction approximately agree with that of modern Earth support a hypothesis of formation by plate tectonic processes, whereas significant differences imply different processes, different rates, or both.

As emphasized in the following sections, continental crust addition and loss are most important today at subduction zones. These include ocean-margin subduction zones (those subducting oceanic lithosphere, e.g. Mariana or Chile) and lithospheric sutures where continents collide (e.g. India with southern Asia and Australia with eastern Indonesia). Our analysis is not limited to plate tectonic processes (i.e. those related to subduction, seafloor spreading, and transform activity); it also considers hot-spot additions and loss by delamination, processes that create and destroy crust independent of whether or not plate tectonics operates. Such processes are likely to have been important on Venus and Mars and should also have been important through much of Earth history.

Our perspective, that continental crust is destroyed as well as created, may be difficult for some to accept. Western geoscientists are imbued by the idea of 'progress' implicit in discussions of crustal 'evolution' (generally taken to be progressive increase in crustal volume), but this perspective restricts broader thinking about the problem. We prefer to distance ourselves from such concepts of linear change and consider a wider range of possibilities for how the volume of continental crust is changing now. In Chinese philosophy, the concept of *yin yang* (simplified Chinese: 阴阳; pinyin: yīnyáng – often referred to in the West as *yin* and *yang*) is used to describe how seemingly opposing forces are inextricably bound together, intertwined, and interdependent in the natural world, giving rise to each other in turn. Many natural dualities – e.g. dark and light, female and male, low and high – are cast in Chinese thought as *yin yang* (Figure 1). The relationship between the two is often described in terms of sunlight playing over a mountain and in the valley. *Yin* (literally the 'shady place' or 'north slope') is the dark area occluded by the mountain's bulk, while *yang* (literally the 'sunny place' or 'south slope') is the brightly lit region. As the sun moves across the sky, *yin* and *yang* gradually and reciprocally change, revealing what was obscured and obscuring what was revealed. *Yin* is usually characterized as slow, soft, insubstantial, diffuse, cold, wet, and tranquil. It is generally associated with the feminine, birth and generation, and with the night. *Yang*, by contrast, is characterized as hard, fast, solid, dry, focussed, hot, and aggressive. It is associated with masculinity and daytime.

The *yin yang* conceptual framework is well-suited for our analysis, because it weds modern processes of crustal formation and destruction, mostly but not only through the subduction process. The *yin yang* analogy is especially appropriate because some of the same tectonic processes that create continental crust – such as subduction – also destroy it. The analogy is also appropriate because some of the processes are so deeply hidden – especially delamination of lower crust and subduction – that they pose special challenges for our understanding and realistic quantification. Important mystery is implicit in *yin yang*. Consequently, the following analysis is not exhaustive – many details are omitted – but we hope to touch on the most important considerations, as we presently understand them. Our discussion

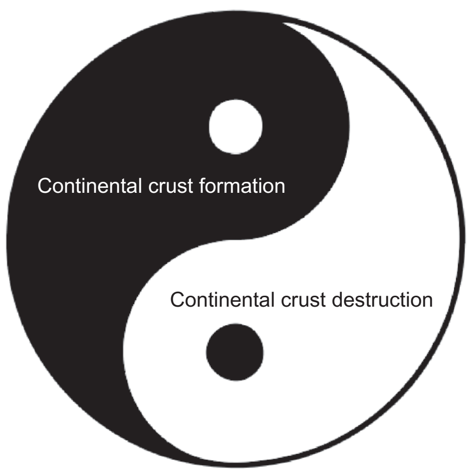

Figure 1. Symbolic summary of the principal conclusions of this article. Plate tectonics forms (yin) continental crust at the same time that it destroys it (yang). Both processes dominantly occur at convergent plate margins.

is intended as a broad exploration of these processes, a 'guide for the perplexed' (including ourselves) and perhaps a roadmap for today's students and future studies.

Plate tectonic styles of continental crust formation and destruction

A linear, net crustal growth rate of approximately 1.75 km^3/year (1.75 Armstrong Units, or AU; Kay and Kay 2008) over the past 4.0 Ga could have built the present approximately 7 billion cubic kilometres of continental crust. Alternatively, this volume could be due to a net, linear crustal destruction rate of approximately 2.6 AU over the same time period, assuming that a primordial, Earth-covering layer of continental crust 40 km thick once existed (Figure 2). Neither of these end-member scenarios are realistic. There is a strong episodicity that demonstrates punctuated evolution of Earth's continental crust (O'Neill *et al.* 2007), including approximately 250 million years (2.45–2.2 Ga) when negligible crustal growth occurred (Condie *et al.* 2009).

In contrast to a broad range of opinions about how the modern inventory of continental crust came to be and what it was in the past, a broad geoscientific consensus exists about

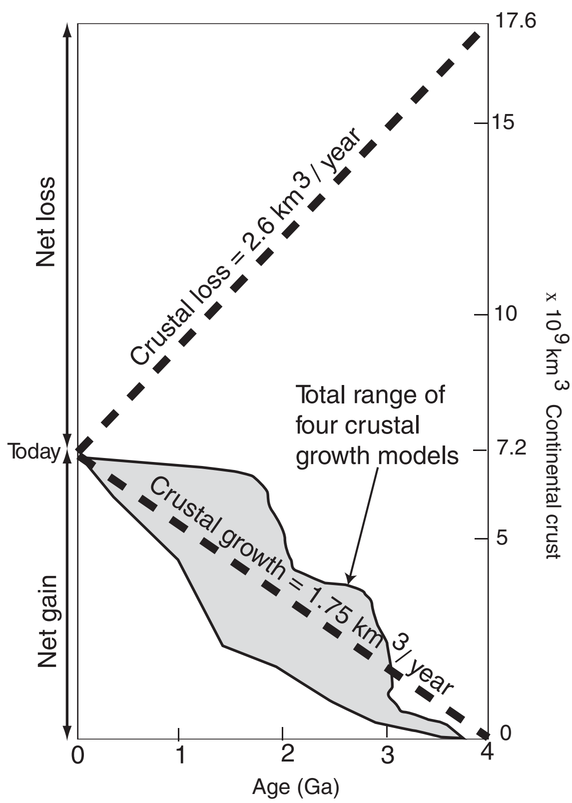

Figure 2. Estimated continental growth rates on a cumulative basis. Dashed lines show that either continuous production of 1.75 km^3/y (lower line) attended by no recycling losses could produce all continental crust that presently exists (7.2×10^9 km^3) or that continuous destruction at a rate of 2.6 km^3/y (upper line) of a primordial continental crust (17.6×10^9 km^3) that once covered the entire Earth could have resulted in the existing volume. Neither of these options is likely to be realistic, but they show the range of possibilities that can explain the present volume of continental crust. Grey field encompasses four crustal growth curves, from Jacobsen (1988). See text for further discussion.

modern processes of crust formation and destruction. Such crust today is mostly generated above subduction zones, with secondary sites associated with rifts, hotspot volcanism, and volcanic rifted margins (Figure 3; Reymer and Schubert 1984; Hawkesworth and Kemp 2006). Subduction zones are also the most important sites of crust removal, by sediment subduction, subduction erosion, and deep subduction of continental crust. Continental crust is also lost by non-plate tectonic processes of delamination of lower crust – which we prefer to call 'foundering' (Figure 3).

Modern processes of crustal loss and production work together to produce juvenile continental crust, to process juvenile crust into true continental crust, and to destroy and recycle both juvenile and mature continental crusts. Losses due to subduction, given enough time, are nearly total, but subduction zone processing of subducted sediment supplements abundances of large ion lithophile (LIL) and light rare earth elements extracted by melting depleted mantle. Foundering of mafic lower crust or subcrustal pyroxenites may be required to yield the appropriate intermediate-composition crust from juvenile, mantle-derived, mafic crust. In this circumstance, crustal refinement is accomplished by fractionation and anatectic melting to yield granodioritic upper crust as well as residual and cumulate gabbroic lower crust accompanied by loss of dense, pyroxene- and garnet-rich subcrustal material. Foundering can lead to new magmatic episodes and thus crust formation (Gao et al. 1998; Bédard 2006; Elkins-Tanton 2007). In spite of the fact that these processes are related, it is useful to separate these processes in order to better articulate them, as done below.

Crust formation

A strong geoscientific consensus exists that continental crust mostly forms today and over the time that plate tectonics has operated by magmatic additions due to water-induced melting of mantle above subduction zones (Coats 1962; Dimalanta *et al.* 2002). This produces juvenile (intra-oceanic) arc crust, such as that of the Izu–Bonin–Mariana arc, and

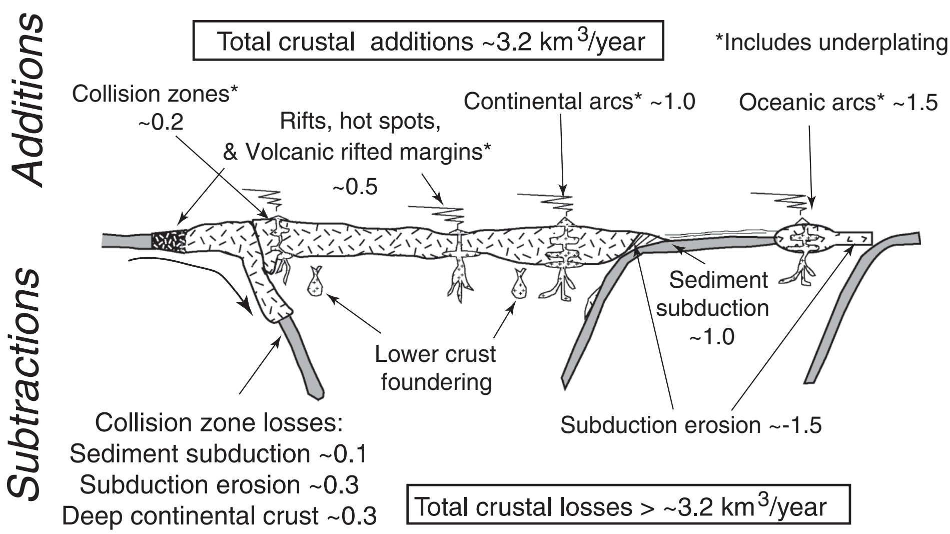

Figure 3. Synopsis of how continental crust forms today, note rates of crustal destruction are more poorly constrained than crustal generation. A maximum crustal production rate of approximately 3.2 km^3/year (or 3.2 AU) is estimated.

magmatically adds to pre-existing continental crust, such as that beneath the Andes. Juvenile continental crust is mostly extracted from the mantle as mafic (basaltic or boninitic) magma, which must be further refined by anatectic remelting of the crust and delamination (foundering) to yield andesitic material approximating the bulk composition of continental crust (Rudnick and Gao 2003; Tatsumi 2005; Tatsumi *et al.* 2008). Although juvenile continental crust is likely to be mafic at first, it inherits from subduction zone processes the distinctive geochemical characteristics of subduction-related magmas: enrichment in silica and LIL elements (e.g. K, U, Sr, and Pb), as well as depletion of high-field-strength incompatible elements (e.g. Nb, Ta, and Ti). These distinctive chemical characteristics persist through later processing by anatexis within the crust and foundering of its base, so that juvenile mafic crust matures into continental crust with andesitic bulk composition (Figure 4). The similarity of trace element variations in arc rocks to that of bulk continental crust strengthens the inference that subduction-related magmatism is an important way to make continental crust today (Rudnick and Gao 2003).

The rate of crust formation above subduction zones may change because of variable forcing functions, such as mantle potential temperature, amount of water, melt, or supercritical fluid added from the subducted slab to the mantle wedge, and the extent of adiabatic decompression of asthenosphere in the mantle wedge. Recent studies indicate that the extent of subduction-related mantle melting (F) increases with increasing convergence rate owing to increased temperatures along the melting front (Cagnioncle 2007). This is because faster subduction recycles more water into the mantle as the sinking slab drives greater asthenospheric circulation in the mantle wedge, which induces more melting and crust formation. Other important controls on melt production rate are the position of isotherms above the wet solidus; this also varies with slab dip (Grove *et al.* 2009). These

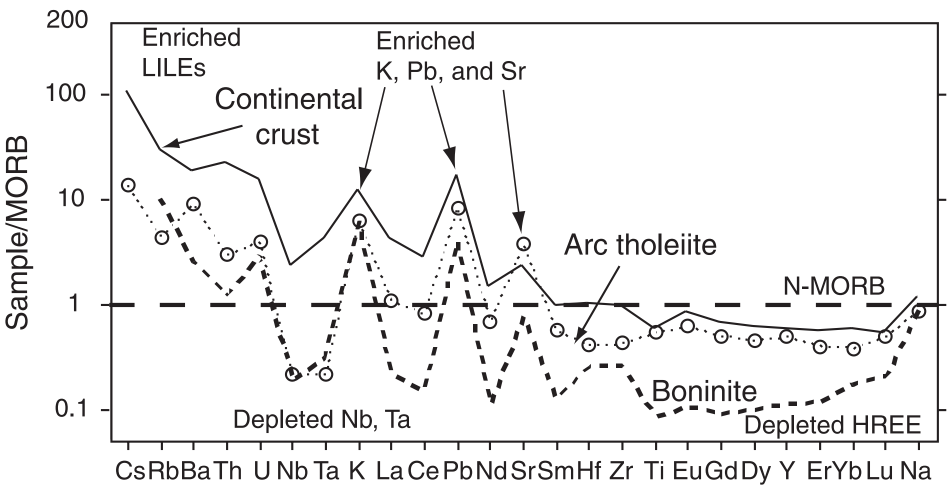

Figure 4. Incompatibility plot (spidergram) for continental crust and juvenile arc melts (boninite, arc tholeiite, and continental crust), modified after Stern (2002). Elements on the horizontal axis are listed in order of their incompatibility in the mantle relative to melt; elements on the left are strongly partitioned into the melt whereas those on the right are strongly partitioned into peridotite minerals. Note the characteristic enrichments of subduction zone outputs relative to MORB with respect to fluid-mobile LIL elements: Rb, Ba, U, K, Pb, and Sr and the relative depletion of these in HFSE: Nb, Ta, Zr, Ti, Y, and HREE. Note also the greater enrichment and overall similarity of continental crust to juvenile arc crust (Boninite, arc and MORB) on the other.

potential forcing functions vary considerably among arc systems worldwide today and surely have over time as well.

Scholl and von Huene (2009) estimate that over the last tens to hundreds of millions of years, approximately 2.7 km^3 of juvenile continental crust has been generated by convergent margin igneous activity each year (Figure 3). This is twice of the rate of Reymer and Schubert (1984), who earlier estimated that arc magmatic activity produces about 30 km^3 of new crust per km every million years (km^3/km-million years), for a global rate of 1.1 AU. These estimates are reconsidered below.

Three independent approaches to calculate arc crustal growth rates need to be considered – (1) geophysical, (2) eruptive volumes, and (3) fossil arcs. The first uses deep-penetrating geophysical surveys to estimate the underlying volume of arc magmatic material imaged along a seismic profile that images as deeply as the Moho, then dividing this volume by the time since the subduction zone (and arc system) began. This approach is best applied to active intra-oceanic arcs (i.e. those formed away from the continents, within the oceanic crustal realm) where we know when subduction began, such as Izu–Bonin–Mariana, Tonga–Kermadec, and the offshore Aleutian Ridge west of the Alaska continental shelf. The second method takes estimates of eruption rates and multiplies extrusive volumes by a factor to approximate the corresponding, hidden plutonic addition. The last approach considers outcrop area and metamorphic P–T facies of fossil arc massifs to infer crustal volumes and couples these with geochronological constraints for the lifespan of the arc to infer crustal addition rates. As explored below, an important – and paradoxical – result of these three approaches is that the first and third approaches yield much larger crustal growth rates than does the second. This is most reasonably explained if magma productivity varies greatly over the life of an arc and is much higher when subduction begins than for mature arcs.

Using the geophysical approach, the most recent estimates for intra-oceanic arc crustal growth rates are 130–135 km^3/km-million years for the Izu–Bonin–Mariana (Takahashi *et al.* 2007) and the Aleutian arcs (Scholl and von Huene 2009). The largest sources of error are as follows: (1) uncertainty about whether to include an original 6–7 km thickness of 'trapped' oceanic crust; (2) where to draw the Moho beneath arcs, which is sometimes inferred above regions with Vp as low as 7.5–7.6 km/s (Fliedner and Klemperer 2000; Takahashi *et al.* 2007; Calvert *et al.* 2008); and (3) how much to correct for loss of arc crust due to subduction erosion. The first uncertainty concerns how arcs form; early concepts assumed that new arcs grew on a substrate of pre-existing oceanic crust that was trapped when the new subduction zone formed (Dickinson and Sealy 1979). More recent considerations indicate that the substrate for intra-oceanic arcs is produced when subduction begins (Stern and Bloomer 1992; Hall *et al.* 2003), as discussed further below. The latter conclusions are especially important for arcs that form spontaneously by lithospheric collapse (Figure 5A; Stern 2004) and may be equally important for forced (induced) nucleation of subduction zones, for example the Aleutians (described by Scholl 2007), although the origin of this arc/subduction zone system is poorly understood.

The second source of error is likely to significantly underestimate crust production, if some of the high-velocity (Vp > 7.5 km/s) lower crust lies below where the Moho is drawn. A couple of geophysical studies using the same data set for the Aleutian arc illustrate this uncertainty. Fliedner and Klemperer (2000) carried out a wide-angle seismic experiment parallel to the central Aleutian arc to determine that the crust was approximately 30 km thick and that the uppermost mantle is quite heterogeneous, with velocities that range from 7.6 to 8.2 km/s. Shillington *et al.* (2004) reinterpreted the same data set to indicate a crust that is 35–37 km thick. They interpreted that regions with high P-wave

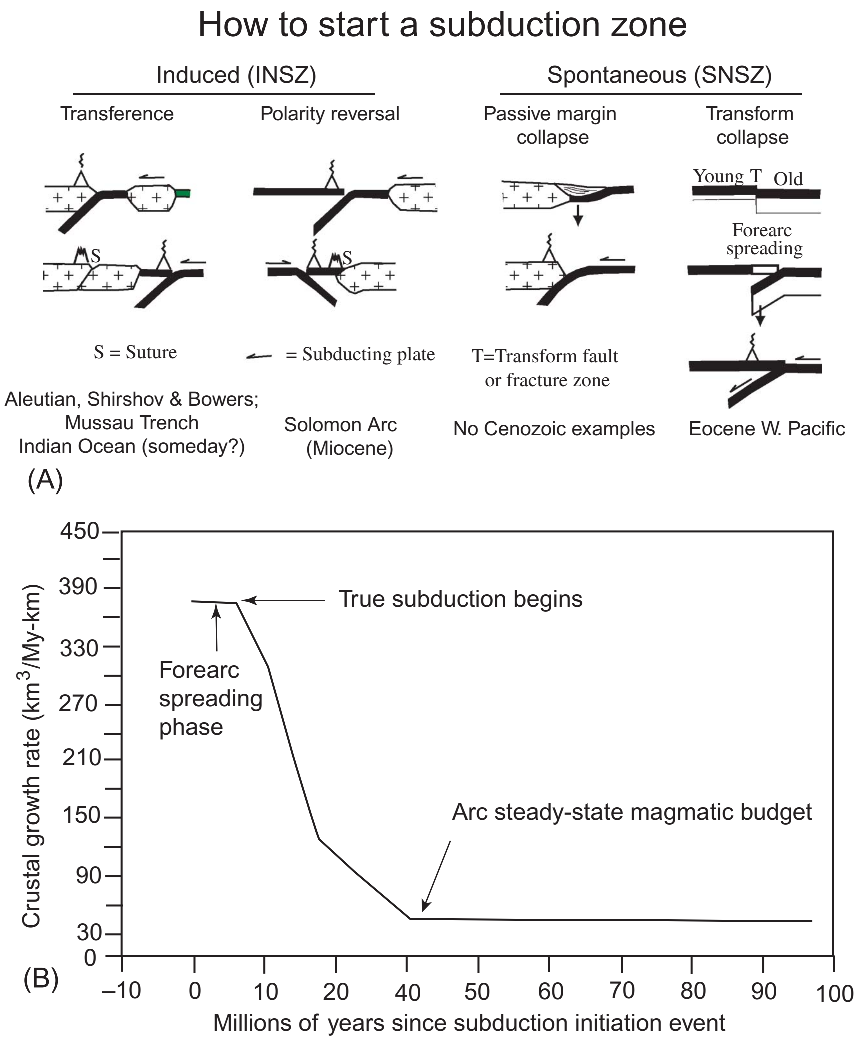

Figure 5. (A) Ways to start new subduction zones (Stern 2004) and (B) effects of subduction initiation mode on magma production. Induced nucleation of subduction zone (INSZ) requires that existing plate motion forces a new subduction zone to form; spontaneous nucleation of subduction zone (SNSZ) occurs before plate motion is directed towards the trench and causes a major change in plate motion. Most Cenozoic examples of INSZ are polarity reversal. All Cenozoic examples of SNSZ are transform collapse. Expected effect of subduction zone initiation mode on magma production rates is intended to be illustrative and is discussed in text.

velocities (7.3–7.7 km/s) to be lower crust composed of ultramafic–mafic cumulates and/or garnet granulites. They also found upper mantle velocities of 7.8–8.1 km/s, significantly higher than that inferred by Fliedner and Klemperer (2000). In both studies, the Moho was identified on the basis of wide-angle reflection results, where the position of reflectors were allowed to 'float' within the velocity model. Whether or not the reflection picked as Moho corresponds to the crust-mantle boundary is debatable, but it is clearly an important lithospheric boundary (Klemperer, pers. com. 2008); in any case, these uncertainties (in the order of 20% of calculated volume) result in a corresponding uncertainty in the calculated crustal growth rate.

The geophysical approach described above applied to the Aleutians and Izu–Bonin–Mariana intra-oceanic arc systems, which have evolved since 45–50 Ma, results in long-term (i.e. over the life of the arc) crustal growth estimates of 90–180 km^3/km-million years (Holbrook *et al.* 1999; Jicha *et al.* 2006; Takahashi *et al.* 2007). This estimate is particularly impressive because it considers only crust above seismic Moho and significant quantities of lower crustal cumulates and residues probably lie below this, as previously mentioned and as discussed further below. This estimate also ignores forearc crust lost due to subduction erosion.

The second way to estimate arc crust magmatic growth is to determine recent eruption rates (readily ascertainable) and multiply this by proportions of extrusive to intrusive igneous rocks. For example, eruption rates are reported for the NE Japan arc by Kimura and Stern (2008). This is a very active volcanic arc, approximately 200 km wide with more than 48 volcanoes that erupted during Quaternary time. Two rows of volcanoes define volcanic front and rear arc chains, which erupt low- to medium-K and high-K suite lavas, respectively (Kimura and Yoshida 2006). NE Japan eruption rates since 2 Ma, including both volcanic front and rear arc, average 2.11 km^3/km-million years. A similar inventory of eruption rates for Central American volcanoes (Nicaragua and Costa Rica) over the past 600,000 years by Carr *et al.* (2007) estimated the volcanic flux at 5–9 km^3/km-million years. Hildreth (2007) estimated eruptive volume for the Quaternary Cascades (2.84 km^3/km-million years), Aleutians (1.21 km^3/km-million years), and Andes SVZ (2.1 km^3/km-million years). Sample and Karig (1982) estimate a volcanic production rate of 12.4 km^3/km-million years for the Mariana arc over the past 5 million years. These six estimates indicate that every million years, arc volcanoes erupt a few to several cubic kilometres of material for every kilometre of arc.

Remarkably, these recent arc eruption rate estimates are almost two orders of magnitude lower than geophysically based estimates of intra-oceanic arc crustal growth. Much of this discrepancy is because eruptive fluxes are only part of the total igneous growth of arcs and the plutonic flux must also be considered. Unfortunately, the volumetric proportions of volcanic to intrusive rocks at convergent plate margins are almost unconstrained. White *et al.* (2006) estimated that the proportions of intrusive and extrusive igneous rocks in arcs range from 1:1 to 16:1, and this is likely to vary with melt composition, crustal thickness, and tectonic setting. Magma composition should influence the intrusive : extrusive ratio, but this control may not be simple. Denser mafic magmas are more likely to pond deeper in the crust than less dense felsic melts (Herzburg *et al.* 1983), yet the higher viscosity of more polymerized felsic melts also favours plutonic emplacement over eruption. Water-rich melts degas and rapidly crystallize as they rise (e.g. Johnson *et al.* 2008), so that drier magmas are more likely to erupt than wetter ones. The composition of the crust also influences the intrusive : extrusive ratio. Felsic crust is less dense, which favours magma ponding in the crust (intrusions) over eruptions and also favours anatexis and fractionation (thus favouring the generation of felsic melts) at the same time that hot lower crust in such realms is more likely to sink back into the mantle.

Tectonic regimes may also exert important controls on the intrusive : extrusive ratio (Glazner 1991), which will be higher where arcs are compressed (and magmas cannot find easy routes to the surface and consequently are more likely to pond in the crust) and lower where they are stretched (and magma easily rises through the crust). These considerations suggest that the intrusive : extrusive ratio will vary over the life of an arc, lower for a young arc with thin, mafic crust and higher for a mature arc with thicker, felsic crust. Regardless of these considerations, assuming an eruption rate of 5 km^3/km-million years and intrusive : extrusive ratio = 10 – which seems high to us – yields a crust production rate that is only about half that of geophysically estimated rates for intra-oceanic arcs. Lower, more reasonable ratios of 3:1 to 4:1 result in crustal production rates that are small fractions of the geophysically estimated rate.

A final perspective on arc crustal growth comes from considering fossil intra-oceanic arc systems, such as the >1000 km-long, Jurassic Talkeetna arc in Alaska, which was active for approximately 35 million years, from about 202 to 167 Ma (Rioux *et al.* 2007). In cross-section, the reconstructed arc is approximately 100 km wide and approximately 30 km thick (Kelemen *et al.* 2003). In addition, approximately 80 km of forearc crust may have been removed between 200 and 160 Ma (Clift *et al.* 2005). The original Talkeetna forearc is known to be gone, because there is now an accretionary/underplated complex juxtaposed against approximately 200 Ma arc gabbros and peridotites, suggesting that the original forearc was tectonically removed by either regional strike-slip faulting (which we think unlikely), subduction erosion, or subaerial erosion to expose the accretionary complex that formerly deeply underlay the igneous massif of the Talkeenta forearc. If the missing forearc crust was 15 km thick, a total of 1200 km^3/km of Talkeetna forearc crust has been removed. As a result we conclude that a total of 4200 km^3/km of Talkeetna arc crust (3000 km^3 of main arc + 1200 km^3 of vanished forearc) was generated during the arc's 35 million years-long history. Thus, the long-term magmatic accretion rate was at least 90 km^3/km-million years and could have been 120 km^3/km-million years, similar to the geophysical estimate for modern intra-ocean arc massifs discussed earlier.

Why the great difference between estimates of arc crustal growth discussed above? The estimates given by the geophysical and fossil-arc volume approaches are much higher than crustal production at mature long-lived arcs unless 20 times as much magma is intruded than erupted, which seems too high. Given an intrusive : extrusive ratio of a few to several, the only reconciliation is if arc magmatic production decreases tremendously over the life of an arc and that recent eruption rates reflect only the low magma production rate of mature arcs. There is strong evidence that arcs grow much faster when they begin. Several studies conclude that magma production above an incipient subduction zone is exceptionally voluminous (Stern and Bloomer 1992; Stern 2004; Jicha *et al.* 2006, Scholl and von Huene 2009). These studies concluded that growth of intra-oceanic arc crust is dominated by a few million years of very rapid crustal growth – perhaps several hundred to as much as 500–600 km^3/km-million years, roughly equivalent to a medium-rate spreading ridge (Figure 5B). Early bursts of crustal growth are followed by much slower growth – perhaps a few tens or less of km^3/km-million years for mature, established arcs, essentially what is inferred from modern eruption rates.

Understanding early voluminous arc volcanism is further complicated because how subduction zones form may also affect early arc crustal growth rates. Stern (2004) identified two general mechanisms for starting new subduction zones: induced (INSZ) and spontaneous nucleation of subduction zone (SNSZ) (Figure 5A). INSZ responds to continuing plate convergence following collision (jamming of a subduction zone by buoyant crust), as illustrated by the examples on left side of Figure 5A. INSZ results in regional compression, uplift, and underthrusting to cause the lithosphere to buckle, yield, and ultimately start a new subduction zone. There are two types of INSZ: transference and polarity reversal. With transference INSZ, the new subduction zone forms seaward of the failed one. This is happening today at the Mussau Trench, may someday occur SE of India, and may have happened in the early Eocene to form the Aleutian subduction zone (Scholl 2007). Polarity reversal INSZ also follows collision, but continued convergence forms a new subduction zone behind (landward of) the magmatic arc; for example, the response of the Solomon convergent margin after colliding with the Ontong Java Plateau (see discussion in

Mann and Taira 2004). Collision (subduction) of the Ontong Java Plateau with the Melanesian arc flipped subduction northward and quickly led to accretion (by obduction) of this arc massif to the northern periphery of the Australia plate and the deformation of Papua New Guinea (Cloos *et al.* 2005).

SNSZ results from gravitational instability of old, dense oceanic lithosphere. Two subclasses of SNSZ are illustrated on the right half of Figure 5A. Lithospheric collapse occurs either at a passive margin or at a transform/fracture zone. The SNSZ hypothesis predicts that seafloor spreading in what will become the forearc predates plate convergence and thus true subduction. Spreading leads to unusually high degrees of mantle melting due to combined effects of adiabatic decompression followed by a large flux of water from the sinking lithosphere. This is where and when most ophiolites form. Transform collapse SNSZ appears to have engendered many new subduction zones west of the Pacific Plate during Eocene time. Stern and Bloomer (1992) argued that the Izu–Bonin–Mariana intra-oceanic arc system in the western Pacific formed by collapse of old Pacific lithosphere east of a major transform, next to much younger seafloor and hotter shallow mantle to the west. Stern and Bloomer (1992) reconstructed the Izu–Bonin–Mariana arc system crust and inferred from this that melts were generated at approximately 120–180 km^3/km-million years during the first 10 million years of subduction. This may significantly underestimate early arc crustal growth. Recent geophysical studies of the Izu–Bonin–Mariana and Aleutian arcs infer mean crustal growth rates during the past 50 million years of approximately 100 km^3/km-million years or more, requiring much higher rates early in the history of these arcs (Dimalanta *et al.* 2002; Jicha *et al.* 2006). Scholl and von Huene (2009) calculate that, during its first approximately 10 million years, Aleutian arc crust grew as fast as 500 km^3/km-million years.

What causes such tremendous early crustal growth, which evolves to longer-lived, lower rates of arc crust production? Early rapid growth reflects wholescale upwelling of asthenosphere up and over the sinking lithosphere (Hall *et al.* 2003), followed by a progressive diminution of upwelling as early slab sinking is replaced by the nearly down-dip motion of the slab, characteristic of true subduction. Development of self-sustaining subduction in the case of SNSZ is signalled by the beginning of down-dip slab motion, which chokes off asthenospheric upwelling, allowing the forearc mantle to cool. As a result, the magmatic arc retreats to a position that is 100–200 km from the trench. These changes cause magma production to decrease greatly following the subduction initiation/arc infancy phase.

It is not yet clear whether INSZ early magma production rates are as great as those for SNSZ. There is no obvious increase in magma production associated with the Solomon Is. INSZ. On the other hand, very large volumes of Aleutian early magma production are inferred to be associated with INSZ (Scholl 2007). The question of what controls magma production when subduction zones form clearly needs more attention. Scholl and von Huene (2009) infer that modern intra-ocean arcs have been building arc massifs since approximately 50 Ma, on average. Accreted arcs, for example the early Jurassic Talkeetna and Bonanza arcs of Alaska, were magmatically active for about this long (Clift *et al.* 2005). Condie (2007) concluded that intra-oceanic arc systems typically exist for <100 million years before they are extinguished by collision. Adopting an average life span of 70 million years, the long-term arc magmatic productivity rate might be approximately 100 km^3/km-million years (500 km^3/million years for first 10 million years plus 30 km^3/million years for 60 million years thereafter). At this rate, crustal growth at the present 14,500 km of intra-oceanic arcs alone (principally the Aleutian, South Sandwich, Kermadec, Tonga, Vanuatu, Solomon, New Britain, Izu–Bonin–Mariana, Indonesia east of Java or Banda, and the Lesser Antilles arcs) would alone average approximately 1.5 AU (Figure 3, Table 1).

Andean-type margins seem to be longer-lived than intra-oceanic arcs, and because of this, less productive over their lives. Scholl and von Huene (2009) note that initial high rates of arc magmatism are progressively muted by the long continued activity. For example, productivity estimates for the Andes since mid-Cretaceous time range from 4 to 35 km^3/km-million years (Kukowski and Oncken 2006) and 35 km^3/km-million years since mid-Miocene time (Haschke and Gunter 2003). The 35 km^3/km-million years rate is essentially the rate of Reymer and Schubert (1984). It implies a reasonable intrusive : extrusive ratio of approximately 7:1 and is adopted as a useful long-term rate estimate for continental margin arc productivity. The combined length of continental margin arcs is approximately 27,500 km (Scholl and von Huene 2007). The corresponding global addition rate is thus approximately 1.0 AU. Combined with the estimates for offshore arcs (1.5 AU), the mean annual growth rate due to igneous activity at convergent margins is thus approximately 2.5 AU. Magmatism also occurs at collisional orogens, but this is mostly crustal reworking. Some igneous activity in collision zones is juvenile, perhaps the 0.2 AU estimated by Scholl and von Huene (2007). Adopting this estimate makes a total of 2.7 AU of crustal growth at the three types of convergent margins (Figure 3, Table 1).

This discussion is certainly incomplete but it shows that in order to understand how juvenile continental crust is generated by plate tectonic processes at modern arcs, we must better understand how subduction zones form and how this impacts early arc magma generation.

Whatever the tectonic environment, formation of continental crust requires two stages: mantle melting to generate basalt and mafic crust and further refinement of this to yield granitic crust and refractory material (Hawkesworth and Kemp 2006). This is accomplished by distilling felsic magmas from mantle-derived juvenile crust via anatexis of amphibolites and fractional crystallization of mafic magmas (Figure 6). Much basaltic

Table 1. A balance sheet for modern continental crust.

Additions (AU)	This study	Clift *et al.* (2009)
Additions at ocean margin subduction zones		
Intra-oceanic arcs	1.5	
Continental arc	1.0	
Total	2.5	2.98
Additions at crust-suturing subduction zones	0.2	
Additions at rifts, hotspots, and volcanic rifted margins	0.5	
Total additions	3.2	2.98
Losses		
Losses at ocean-margin subduction zones		
Sediment subduction	−1.1	1.65[a]
Subduction erosion	−1.4	1.33[a]
Total	−2.5	2.98[a]
Losses at crust-suturing subduction zones		
Sediment subduction	−0.1	
Subduction erosion	−0.3	
Continental crust	−0.3	0.4[b]
Losses due to foundering	Unknown	
Total losses	>3.2	3.38

Data from Clift *et al.* (2009) are for purposes of comparison with our estimates.
[a]Indicates total for all convergent plate margins.
[b]Indicates subduction of passive margins.

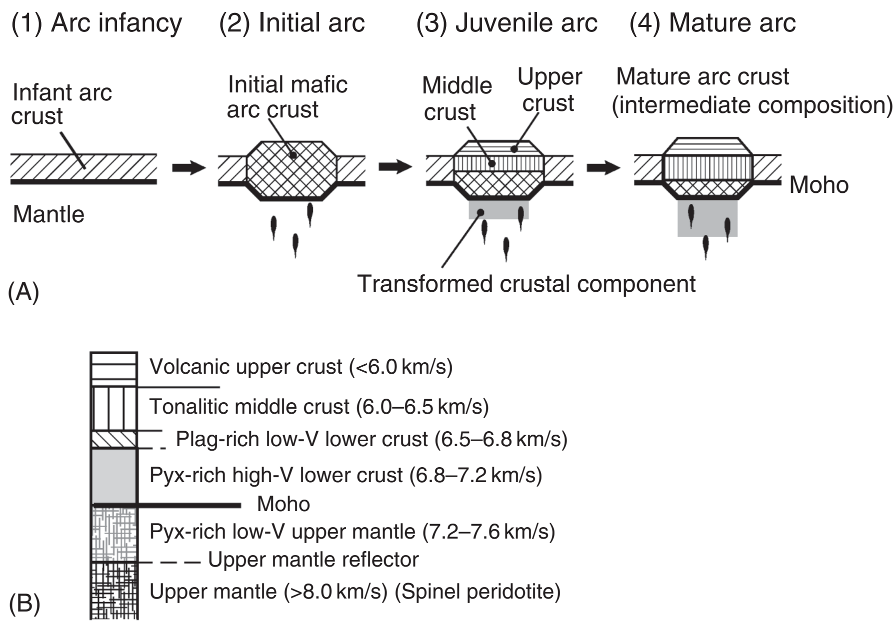

Figure 6. Formation and refinement of juvenile arc crust, like that of Izu–Bonin–Mariana. (A) Model of arc crust evolution from Tatsumi *et al.* (2008). Incipient arc magmatism replaces the pre-existing oceanic crust to create the initial mafic arc crust. Continuing arc magmatism causes anatexis and differentiation of the arc crust, along with transformation of mafic crustal component into the mantle through the Moho, finally creating mature arc crust with an intermediate composition similar to the average continental crust. (B) Generalized seismic velocity structure in the Izu–Bonin–Mariana arc after Suyehiro *et al.* (1996), Takahashi *et al.* (2007), and Kodaira *et al.* (2007).

magma crystallizes as pyroxene-rich material near the Moho. Juvenile arc crust also contains significant admixtures of igneous rocks that are enriched in silica relative to basalt such as boninite and high-Mg andesite (Kelemen 1995) and adakite (Defant and Drummond 1990). Many adakites are thought to be melts of subducted oceanic crust, but these are minor components of modern magmatic arcs (Richards and Kerrich 2007). Adakite-like melts can also result from fractionation of basaltic magma at pressures high enough to stabilize garnet. Adakitic melts may be generated by anatexis at the base of thickened continental crust, where garnet is stable, but this is recycled or reprocessed continental crust, not new additions. Richards and Kerrich (2007) summarized several examples of rocks with adakite-like geochemical signatures that may have formed by partial melting of lower crustal garnet amphibolites or eclogites, from Tibet, the Philippines, the Cascades, southern Sierra Nevada, Baja California, and New Zealand. Adakitic rocks that do not result from melting of subducted lithosphere are expected to form wherever garnet, amphibole, or titanite fractionation is important in the magmatic evolution of mafic melts. Presumably, similar processes beneath hot spots, continental rifts, and volcanic rifted margins could also yield adakite-like melts.

Additions to the continental crust are also made at volcanic rifted margins that form when continents rift apart (Figure 3). Break-up is often associated with eruption of thick piles of mostly mafic lava at the continent–ocean crustal boundary (Menzies *et al.* 2002; Mjelde *et al.* 2007). This mafic crust requires further processing as outlined in Figure 6. Subordinate volumes of felsic material erupted at volcanic rifted margins demonstrate that new continental crust is added here (Bryan *et al.* 2002). Bryan *et al.* (2002) infer that partial melting of amphibolitic crust generates most volcanic rifted margin felsic magmas. Because the amphibolite is derived from volcanic rifted margin mafic magmas at depth, these felsic melts are juvenile additions to the continental crust.

New continental crust also can form by non-plate tectonic processes such as hotspot igneous activity (Figure 3). Such magmas are dominated by tholeiitic basalts that are added to the crust by underplating and intrusion, and these mafic solids may be further processed by anatexis and fractionation to yield felsic melts and foundering of residues. Such processes may have occurred beneath the eastern Snake River Plain recently and may be happening today beneath the Yellowstone hotspot (e.g. McCurry *et al.* 2008). Juvenile crustal growth can also happen by accretion of oceanic plateaux (hotspot products) to arcs and continental crust, as recently occurred where the Ontong–Java Plateau collided with the Solomon arc (Mann and Taira 2004). Accretion of oceanic hotspot constructions such as plateaux and seamount chains is thought to have contributed greatly to Mesozoic crust formation in the Canadian Cordillera (Lapierre *et al.* 2003), growth of coastal promontories of Kamchatka (Portnyagin *et al.* 2008) and elsewhere in the circum-Pacific region. The relative importance of arc and hotspot magmas for growth of juvenile continental crust is reflected by the models of Hawkesworth and Kemp (2006), who infer a mixture of 8% hotspot and 92% arc for the formation of new continental crust on average.

Destruction of continental crust

Continental crust is destroyed in four principal ways: erosion and sediment subduction, continent subduction, subduction erosion, and lower crustal foundering (Figure 3). The first preferentially removes upper crust, the second and third remove upper and lower continental crust equally, and the last process reduces lower crust. The first three mechanisms require subduction and thus plate tectonics but not the last one. These four mechanisms are discussed further below.

Erosion and sediment subduction

Erosion (and weathering) of the continental crust attacks the upper surface of continents, and this material is transferred to the ocean basins, where it is deposited and often ultimately subducted. This mode of crustal destruction along with subduction erosion has received a lot of recent attention (von Huene and Scholl 1991; Clift and Vannucchi 2004; von Huene *et al.* 2004; Scholl and von Huene 2007; Scholl and von Huene 2009).

There are two types of subducted sediment: pelagic (deposited slowly on the seafloor, far away from the trench) and terrigenous (continent- and arc-derived sediments deposited rapidly in and near the trench). Pelagic sediment comprises biogenic materials, wind-borne clays, and distal volcanic ash. This is largely derived by erosion or dissolution of continental crust or overlying sediments (which formed during earlier episodes of continental erosion or weathering). Wind removes dust from the upper continental crust and ash is ejected from stratovolcanoes built on continents and tapping melts derived from the continental crust. Biogenic sediments (chert, chalk, and limestone) are composed of ions (Ca, Si, and associated trace elements) that were released from the upper continental crust by weathering and river run-off, as well as seafloor hydrothermal activity. Some of these ions are also taken up by oceanic crust as a result of submarine hydrothermal activity and seafloor weathering. Near-trench sediment and that of adjacent abyssal plains are overwhelmingly clastic or volcaniclastic, mostly eroded or ejected from the upper continental crust and delivered by glaciers and rivers to the ocean, where they are subducted. Much of the subducted K and other 'fluid-mobile' large-ion lithophile elements are stripped when subducted sediments dehydrate or melt and are returned to the upper plate crust as arc magmas. Nevertheless, most subducted sediment is lost to the deeper mantle, an inference based on the relatively small volume of deeply subducted sedimentary material preserved in the pre-Mesozoic rock record.

Sediment subduction helps protect the upper plate from tectonic erosion by the subducting lower plate. This is because high sediment flux builds a frontal prism of accreted lower plate sediment that shields the seaward edge of forearc basement from impacting seamounts and ridges (von Huene and Ranero 2009). Thick subducted sediments also charge the subduction channel separating the two plates, promoting sub-forearc underplating and crustal thickening, but may also facilitate crustal losses (as discussed in 'Subduction erosion'). Sediment subduction averages approximately 15 km^3/km-million years of trench at accreting (i.e. widening seaward) sediment-dominated margins. The average sediment subduction rate is estimated as twice this (approximately 30 km^3/million years-km) at sediment-starved or landward-narrowing margins (Scholl and von Huene 2007, Table 2). This seeming incongruity is because large accretionary bodies at sediment-dominated margins sequester approximately 70% of all sediments entering subduction zones. Furthermore, accretionary margins are generally located where the orthogonal underthrusting rate is 35–45 km/million years or lower. At non-accreting margins, 100% of incoming sediments are consumed and the typical underthrusting rate is approximately 80 km/million years (Clift and Vanucchi 2004; von Huene *et al.* 2004; Scholl and von Huene 2007).

Subduction erosion

Subduction erosion removes continental and other upper plate materials as a result of the mechanical and fluid-driven disintegration of the lower landward trench slope caused by the underthrusting bathymetric relief of seamounts, aseismic and spreading ridges, and trains of horst and graben created by the bending of the plate as these enter the subduction zone (Collot *et al.* 2008). Subduction erosion is most important for subduction zones with little sediment and rapid convergence (Clift and Vanucchi 2004; von Huene *et al.* 2004; Scholl and von Huene 2007, 2009). However, although little is factually known about subduction erosion in sediment-charged subduction zones, inferences have been drawn that it does occur, especially where great megathrust ruptures occur, such as within the sediment-nourished Cascadia (1700, Mw 9.0) and south central Chile (1960, Mw 9.4) subduction zones (Wells *et al.* 2003).

von Huene *et al.* (2004) noted that the trenchmost sector of the forearc is a dynamic wedge with elevated pore-fluid pressure caused by the fluids squeezed out of subducted seafloor. Overpressured fluids invade and fracture the upper plate, in concert with weakening of forearc mantle by serpentinization. Fractured forearc rock readily collapses into the trench, where it forms a frontal prism that dynamically is an actively deforming Coulomb wedge of accumulated and tectonized debris. The base of this prism is continuously dragged into the subduction channel at the plate interface (Cloos and Shreve 1988). Landward of the frontal prism, older forearc material is thinned by basal erosion and thickly filled forearc basins form along many convergent margins (Scholl *et al.* 1980; Clift *et al.* 2003; Wells *et al.* 2003; Scholl and von Huene 2007, 2009). Exhumation removes the upper crustal rocks of a former forearc to expose the deeper underplate. This is happening now around the north Pacific rim, for example at Kodiak Island, Alaska (Sample and Moore 1987; Moore *et al.* 1991; Clendenen *et al.* 2003). Basal and frontal tectonic erosion and also surface erosion do most of the forearc demolishing and fore-shortening (Grove *et al.* 2008).

All of the material that is removed from the forearc enters the subduction zone via the subduction channel above the down-going oceanic plate and is transported downward. Where large volumes of subducted sediment are involved, material can return towards the surface by counterflow in the subduction channel, but much of the crust and sediment that enters the subduction zone is carried to great depth and lost to the mantle.

Flat-slab subduction is especially effective at stripping away lower continental crust by basal subduction erosion and, in the example of the Laramide California margin, replaced the lower crust with underplated sediment (Saleeby 2003). This underplate is widely exposed around the north Pacific as accretionary complexes emplaced at depths of 20–40 km and at $T \sim 350°C$.

Where two continental bodies come in contact at a crust-suturing subduction zone the preserved forearc of the upper plate underthrust by the lower plate will have lost much of its original mafic lower crust as a result of basal tectonic erosion. So, evidently both flat-slab under-running and collisional suturing can help make the continental crust more felsic by selectively removing mafic lower crust and replacing it with high P/T sediment.

Scholl and von Huene (2009) conclude that ~60 km^3/km-million years of continental crust is recycled into the mantle by subduction, ~25 km^3 subducted sediment and ~35 km^3 tectonically eroded from the forearc. The corresponding global loss for the present ~43,000 km-length of convergent margins is thus ~2.5 AU. Scholl and von Huene (2009) infer that crust has been removed at an estimated average global length of collisional margins of ~6000 km. This length is based on the estimate of Hildebrand and Bowring (1999) that over geologic time the length of global active sutures has ranged from 1000 to 20,000 km. Presently, the global length of major ongoing suturing is ~12,000 km (today mostly associated with closing of the Tethys and Australia against Indonesia). Eventual closing of the Atlantic to construct a new supercontinent would form a suture at least 13,000 in length. Building the new 'Pangea' will thus entail 25,000 km of crustal suturing. In contrast, there is likely negligible suturing immediately after breakup of a supercontinent and dispersal of these fragments (Figure 7). We thus accept the long-term average of 6000 km-long for continental suturing (Scholl and von Huene 2009) to factor in the likely variation of suture length over a supercontinent cycle. Crustal recycling at suture zones is estimated at ~115 km^3/km-million years. This includes ~15 km^3 due to sediment subduction, 50 km^3 of tectonically eroded upper plate, and 50 km^3 for the loss of deeply subducted lower plate crust, resulting in an additional ~0.7 AU of crustal losses at suture zones. Scholl and von Huene (2009) thus assess the long-term rate of crustal destruction by all the aforementioned processes at ~3.2 AU (Table 1, Figure 3).

Continent subduction

Continental crust is also destroyed by being deeply subducted as part of the downgoing plate during collision. Collision occurs when buoyant crust enters a subduction zone and disrupts its normal operation (Cloos 1993). A good example is the continuing descent of India beneath Asia to create the Himalayas and the Tibetan Plateau, which is also forcing large parts of Asia to escape to the northeast and southeast. Researchers agree that a large amount of Indian crust has been underthrust beneath Asia (e.g. 350–800 km, Harrison *et al.* 1992; approximately 700 km, DeCelles *et al.* 2002), but the ultimate fate of this material is unknown. Similar conclusions that continental crust is deeply subducted result from geologic studies of the Papua-New Guinea collision zone (Cloos *et al.* 2005). Another example

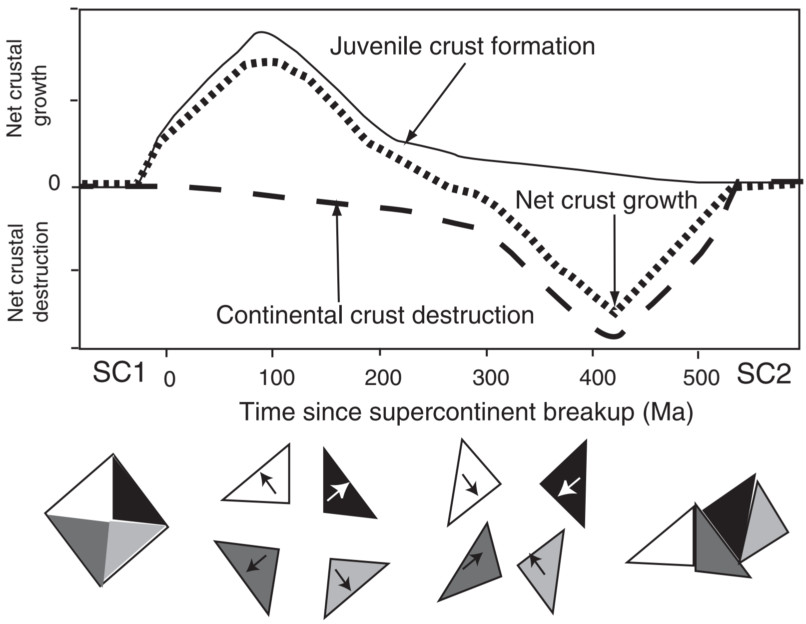

Figure 7. Variations in net crustal growth and destruction over a typical supercontinent cycle. Note that production of juvenile crust is likely to be high following continental break-up (hot-spot activity and volcanic rifted margin formation; development of new subduction zones and arcs) and crustal destruction is likely to be high during continental assembly (higher sediment flux to trenches, more continental crust subducted).

is the continuing subduction of Arabia beneath Eurasia, whereby Arabia continues advancing north at 2–3 cm/year (Vernant *et al.* 2004; Reilinger *et al.* 2006). Convergence of NE Africa, which split off as Arabia approximately 25 Ma, has been continuous since at least 56 Ma even though contact between Arabian and Eurasian continental crust has occurred only in the last 10–20 Ma (McQuarrie *et al.* 2003). Some of this convergence has been accommodated by subduction of East Arabia beneath the Zagros fold-and-thrust belt of Iran.

Conventional geoscientific thought holds that continental crust is too buoyant to be deeply subducted, but geophysical and geological evidence controverts this. The continuing descent of Tethyan oceanic lithosphere must be dragging Indian continental crust down with it. P-wave tomography identifies high-velocity zones beneath the Himalayan collision zone that can be traced to >1000 km deep in the mantle; these must be still-descending slabs of subducted Tethys oceanic and Indian continental lithosphere (Van der Voo *et al.* 1999). We also have direct evidence that at least some Indian continental crust is deeply subducted, because ultra-high pressure (UHP) metamorphic terranes – which contain metamorphic coesite, diamond, unusual Si-rich garnets, and/or K-bearing pyroxenes – are known from near Nanga Parbat and Tso Morai in the Himalayas (O'Brien *et al.* 2001; Kaneko *et al.* 2003; Leech *et al.* 2005). UHP minerals demonstrate that these continental fragments descended to at least 90–140 km depth (Ernst 2006). The metamorphic relations preserved in these rocks and their ages indicate that Indian continental crust was subducted at least 100 km deep within 7–9 million years after India–Asia collision began approximately 50 Ma and then returned to the surface. How much Indian crust has been so deeply subducted and what proportion returns to the surface versus that lost to the deeper mantle is not known. Nor do we know the extent to which lower crust is lost relative to upper crust. Johnson (2002) concluded that only mid/upper Indian crust is found in Himalayan thrust sheets, further inferring that Indian lower crust and mantle detached along zones of weakness or other rheological boundaries (e.g. the Main Central Thrust) and were lost by subduction. The important point is that vast expanses of Indian continental crust may have been lost, that similar losses of continental crust due to deep subduction may be occurring beneath Iran and the Alps, and that other episodes of massive loss of continental crust have happened during those periods that plate tectonics has operated (Armstrong 1991; Bowring and Housh 1995; Hildebrand and Bowring 1999).

Global UHP metamorphic terranes demonstrate that deep subduction of continental crust during collision is common. UHP terranes are known from the Dabie–Sulu belt of China, the western Alps, the Kokchetav Massif of Kazakhstan, the Bohemian Massif of Europe, the Western Gneiss Region of Norway, and the Himalaya. But UHP terranes are often relatively small and thin, and the bulk of the approximately 40 km-thick crust that must have originally existed is missing. UHP complexes typically comprise ductilely deformed thrust sheets that were somehow sliced away from the rest of the descending crust. UHP nappes may be exhumed because of regional mantle upwelling, subduction channel backflow, or underplating combined with extensional collapse (Ernst 2007). Exhumation of majoritic (garnet-pyroxene solid solution) peridotites indicates that material subducted to >200 km depth may be returned to the surface (Liou *et al.* 2007; Scambelluri *et al.* 2008). The return of deeply subducted material to the surface often does not occur, as shown by the fact that no UHP terranes have been identified in North America despite many collisions over geologic time and an abundance of metamorphic petrologists hungry to make the discovery.

The question remains: how much of the continental crust is lost to the deep mantle by subduction? Presently about 20% of Earth's convergent margins are broadly collisional. Oceanic crust is approximately 15% as thick as continental crust, so that the volumetric flux of oceanic and continental crust into subduction zones is subequal today. Significant losses of continental crust in collision zones seem inevitable, but how can we estimate this flux?

One approach is offered by Hacker (2008), who assumed that each of the approximately 20 known UHP terranes was once as large as the best known Norwegian and eastern China examples (~10,000 $km^2 \times 10$ km thick). He used this to calculate a deep flux of subducted continental crust of 200 km^3/million years. Hacker (2008) noted that this is a minimum flux because of incomplete exhumation and opined that the real flux might be >1000 km^3/million years. This estimate assumes that all subducted continental crust eventually returns to the surface, an assumption that we challenge. Especially pertinent in this regard is Ernst's (2006) observation that most UHP terranes are thin sheets, suggesting that much thicker, volumetrically dominant tracts of continental crust are subducted beyond the 'point of no return'.

Another point to consider is that subduction of continental crust is not a steady-state process. Over geologic time, continental collision and loss of continental crust by deep subduction occurs intermittently, for example when supercontinents assemble. It is quite difficult to constrain this flux, but here we attempt to make an estimate: The present area of India is approximately 3,000,000 km^2. Assuming that the India crust is 40 km thick yields a volume of 120,000,000 km^3. As a minimum estimate, let us further assume that 10% of this has been lost by deep subduction since collision began approximately 50 million years ago. 12,000,000 km^3 of continental crust lost every 50 million years averages 0.24 km^3/year, about 8% of the flux (2.9 km^3/y) due to sediment subduction and subduction erosion combined at ocean-margin and crust-suturing subduction zones. The assumed loss of India crust over 50 million years is smaller than but consistent with the estimate of Scholl and von Huene (2009) that the long-term loss of continental crust (i.e. sediment subduction + subduction erosion + detached and foundered lower plate continental crust, Figure 3, Table 1) at all continental collision zones is approximately 0.7 km^3/year. Such estimates do not inspire confidence that we really understand this flux but usefully illustrate another important – and often overlooked – way that continental crust is destroyed.

Lower crust foundering

Lower crust foundering may be the most important way to lose continental crust without subduction. This is likely to be an important process on other silicate planets that are tectonically and magmatically active but do not have plate tectonics, such as Venus (Smrekar and Stofan 1997) and perhaps Mars. A wide range of processes whereby lower crust or lithosphere sinks into the mantle are commonly referred to as delamination (Gögüs and Pysklywec 2008). The term implies peeling off a layer and sinking as a coherent sheet. True delamination is probably uncommon because the viscous resistance of the asthenosphere will impede the detachment and sinking of a 3-D sheet. It is more likely that lower crust is lost by ductile flow-and-drip, first flowing horizontally to create a growing boundary perturbation and then sinking as a gravitational instability. Elkins-Tanton (2007) noted that this mechanism requires no specific structural weakness, only gravitationally unstable lower crustal or mantle lithosphere that is weak enough to flow. Such flow is especially likely where dense mafic lower crust is hot and therefore weak (Jackson 2002); weakness in this layer could also help detach the subjacent lithospheric mantle. A gravitational instability nucleates and grows by lateral or radial flow, allowing mass to gather, enlarging the Rayleigh–Taylor instability. The instability drips into the underlying mantle material and sinks into the mantle as a negative (dense) diapir.

A good example of dripping lower crust was presented by Zandt *et al.* (2004), who used geophysical techniques to infer the removal of a dense root beneath the southern Sierra Nevada batholith of California. Petrologic evidence for foundering in this region was presented by Saleeby *et al.* (2003), who noted a pronounced change in xenolith suites sampled by Pliocene-Quaternary lavas in the southern Sierra Nevada, from early eclogitic (lower crust and lithospheric mantle) to a younger suite of garnet-absent, spinel, and plagioclase peridotites, which they interpreted as asthenospheric samples. They interpreted this change to indicate that much of the sub-Sierran lithosphere was removed in late Miocene to Pliocene time. This flow is presently developed as pronounced asymmetric flow into a mantle downwelling (drip) beneath the adjacent Great Valley. Today, a nearly horizontal shear zone separates the flowing lithospheric root from the overlying batholithic crust. At the top of the drip, a stalagtite-like cone of crust is dragged down tens of kilometres by the mantle drip, erasing the Moho in seismic images.

Lower crustal foundering is especially likely where magmatism creates thick lower crust that is dense, hot, and weak; such conditions exist beneath magmatic arcs and hot spots. The crust-mantle boundary beneath active magmatic arcs is often not well-defined seismically, suggesting that it is chemically transparent as well, with material moving both up and down across it. In these situations, melts rise from the mantle into the crust at the same time that dense mafic lower crust and lithospheric mantle sink back into the circulating asthenosphere (Tatsumi *et al.* 2008). With this *yin–yang* mechanism in mind and not precluding the loss of entire layers by coherent collapse, lower crustal and subcrustal losses due to gravitational instability are more inclusively termed ‘lithospheric foundering’ or ‘lower crustal foundering’, depending on what is lost. This terminology is adopted here.

Foundering material is dense, likely dominated by pyroxene-, garnet-, and/or amphibole-rich cumulates and restites. It is often not clear whether such plagioclase-poor material should be considered as lower crust or upper mantle. Certainly, as a magmatic rock it is petrogenetically linked to the crust but it is often ultramafic in composition and on a seismic basis would be assigned to the mantle. Removal of this material may be the key process in converting mafic juvenile crust into true continental crust, as discussed previously. Müntener and Ulmer (2006) conducted experiments on H_2O-undersaturated, picrobasaltic to andesitic magmas at pressures exceeding 0.8 GPa (~25 km) corresponding to the base of typical intra-oceanic arc crust. They found that 40–60% of ultramafic cumulates are produced by fractionation during formation of the andesitic to dacitic compositions that are typical for evolved arc magmas. Calculated P-wave velocities for these cumulates range from 7.3 to 8.0 km/s. Such material will form a broad crust–mantle transition, if these cumulates do not founder very deeply into the mantle (Figure 6A,B).

There is observational support for the idea that foundering of dense cumulates and restites is important in the evolution of active magmatic arcs. Behn and Kelemen (2006) concluded that lower crust with $Vp > 7.4$ km/s beneath active magmatic arcs is generally gravitationally unstable relative to the underlying asthenosphere and likely to founder. Lower crust below most modern arcs has $Vp < 7.4$ km/s, implying that gravitationally unstable material is generally lost and that dense material founders almost as quickly as it forms. Recent P-wave velocity measurements for the crust and upper mantle of the Izu–Bonin–Mariana arc system (Takahashi *et al.* 2007, 2008; Calvert *et al.* 2008) show that Izu–Bonin–Mariana middle ($Vp < 6$ km/s) and lower ($Vp <7$ km/s) crust is composed of intermediate to felsic and mafic rocks, respectively. Petrologic modelling to explain the composition of arc lavas indicates that more lower crust – composed of olivine- and pyroxene-rich cumulates – should have been produced than is observed from geophysical imaging. Tatsumi *et al.* (2008) concluded that on average 160% as much subcrustal pyroxene-rich materials as juvenile crust underlies intra-oceanic arcs (Figure 6). Much of this seems to have disappeared, most likely by foundering into the upper mantle.

Tatsumi *et al.* (2008) concluded that middle Izu–Bonin–Mariana arc crust of intermediate composition was generated from mantle-derived basalts via complex multistage magmatic processes. These include anatexis of amphibolite-facies mafic lower crust to generate felsic melts and mixing of felsic melts with mantle-derived basaltic magma (Figure 6). They further inferred that the volume of mafic restite and cumulates that are 'crustal residues' resulting from formation of middle and upper arc crust should be three to nine times greater than that of the seismically defined Izu–Bonin–Mariana lower crust. Tatsumi *et al.* (2008) further suggested that the mafic to ultramafic crustal components are transferred to the subarc mantle. During this process, the lower crust and upper mantle is hot and weak enough to allow crustal residues to sink, as well as magmas to rise through the sub-arc Moho. As noted above, seismic velocities beneath the Moho of active arcs are typically low, and perhaps where we draw the Moho beneath arcs needs to be reconsidered. In any case, the separation of dense residues from primitive arc crust seems to be essential for creating compositions that approximate continental crust from mantle-derived basaltic melts.

Fossil arcs give us a chance to directly examine subarc lower crust and upper mantle. An excellent example of a largely intact, fossil intra-oceanic arc showing these relations is found in the Talkeetna Mountains of southern Alaska. Jurassic Talkeetna volcanic rocks are compositionally diverse, ranging from basalt to rhyolite that evolved by fractional crystallization and anatexis; complementary cumulates are preserved as lower crustal gabbronorites and upper mantle pyroxenites. Primary, mantle-derived mafic melts fractionated to form cumulate and residual amphibolite at the base of the crust (DeBari and Sleep 1991). Greene *et al.* (2006) calculated that more than 25 wt% of Talkeetna primary melts crystallized as pyroxenite at the base of the crust but were mostly lost, probably together with dense garnet granulites, foundering into the underlying mantle. Similar observations have been made for the Cretaceous Kohistan intra-oceanic arc of the Pakistani Himalaya, where the lack of cogenetic ultramafic cumulates complementary to the evolved mafic plutonic rocks led Garrido *et al.* (2007) to conclude that either such cumulates formed but were lost from the base of the arc crust or that the net flux to the Kohistan arc crust was more evolved than primitive basalt.

Subcrustal losses beneath intra-oceanic arcs seem to be controlled by the density of olivine-, pyroxene-, and amphibole-rich material, but garnet may also be important beneath the thicker crust of Andean-type arcs. Crustal refinement at Andean-type margins is also accomplished by fractionation and anatectic melting to yield granodioritic upper crust, as well as residual and cumulate gabbroic lower crust and garnet-pyroxenite subcrust (Ducea 2002). In the example of the Mesozoic Andean-type arc represented by the Sierra Nevada of California, Ducea (2002) concluded that granitoids resulted from fractional crystallization of parent mafic melts (with 20–30% anatectic crustal component) at depths exceeding 35–40 km; at these pressures, cumulates and residues are very dense eclogitic (garnet pyroxenite) rocks. Ducea (2002) estimated that the mass of the residual assemblage was one to two times the mass of the granitic batholith. Such dense garnet pyroxenites are prone to founder into the underlying asthenosphere. Ducea (2002) inferred that the average rate of foundering over approximately 120 million years life of the Sierran arc was 25–40 km^3/million years-km.

A final perspective on the importance of foundering comes from recent geochemical considerations. Plank (2005) argued that loss of mafic lower crust is essential for generating true continental crust from juvenile mafic crust. She estimated from global Th/La variations that foundering of 25–65% of cumulates and restite into the mantle can create a bulk continental crust that evolves to high Th/La (>0.2) values that are typical of true continental crust. The loss of such dense material would also serve to increase silica and LIL element contents in remaining buoyant crust.

It is clear from the foregoing discussion that foundering of dense cumulates and restites is critical for forming continental crust at convergent margins. It seems likely that sinking of dense cumulates and residues will occur at all sites of juvenile crust formation, beneath rifts, hotspots, and volcanic rifted margins as well as beneath arcs. Taking Plank's (2005) estimate that foundering occurs at a rate of 25–65% of modern magmatic crust production (=2.7 km^3/y, Figure 3), this implies a foundering rate of 0.7–1.8 km^3/year. But this is not really loss of continental crust, it is an addition or return flow to the mantle during processing of juvenile crust required to form continental crust.

It should be finally noted that lower crustal and lithospheric mantle foundering is likely to happen in a wide range of active tectonic environments, from when the growing intra-oceanic arc begins to thicken (Jull and Kelemen 2001) to when continent–continent collision occurs (Anderson 2005). Similarly, refinement of the crust continues as long as the system is magmatically active; mafic magmas injected at the base of or into the crust will spur further differentiation via anatexis (Annen *et al.* 2006). The Moho beneath active zones of crustal growth must be open to a vertical *yin–yang* mass transfer, with mafic melts being added to the crust at the same time that dense cumulates and residues sink back into the mantle, as discussed above. Processes related to magmatic activity – mafic addition, anatectic refinement, and foundering of crustal residues – appear to be quasi-continuous with igneous activity. These processes are likely to be important wherever significant volumes of mafic, juvenile crust are created, not just beneath intra-oceanic arcs but also beneath rifts, hot-spots, and volcanic margins.

Accretion and orogeny

As discussed above, modern juvenile crust is mafic and progressively becomes more felsic as it develops compositional layering (mafic at the base, felsic at the top). This progression occurs as it is processed from juvenile arc, volcanic rifted margin, or oceanic plateau into stable continental lithosphere, or craton. This evolution can also occur in tandem with terrane accretion, whereby various juvenile crustal fragments coalesce to build progressively larger tracts of increasingly differentiated crust. Accretion accompanies collision, when a subduction zone is choked with buoyant crust and lithosphere (Cloos 1993), destroying the original subduction zone and perhaps forcing a new one to form elsewhere (Stern 2004). Collision involves both crustal gains (accretion of juvenile arcs and plateaus) and losses (erosion and sediment subduction, continental subduction, and foundering), as discussed separately above. Lithospheric and lower crustal foundering as well as the production of anatectic melts are both likely to be enhanced in these hot realms during the thickening and deformation that accompanies terrane accretion. In fact, many orogenies reflect such episodes of terrane accretion, which typically last 8–10 million years (Clift *et al.* 2009).

Summary of crustal growth and destruction today

Annual additions to the crust total 3.2 AU, while crustal losses at ocean-margin subduction zones are at least 2.5 AU (Figure 3, Table 1). This implies that the inventory of continental crust could be increasing by as much as 0.7 AU annually. But, although far less securely estimated, losses of this amount are estimated to accompany the suturing of continental crustal blocks (Figure 3; Scholl and von Huene 2009). There are uncertainties for all of aspects of these estimates but the greatest uncertainty surrounds losses by continental subduction and by losses of lower crust by foundering. If these crustal losses combined are greater than 0.7 AU, then modern tectonic processes decrease the inventory of continental crust with time, a possibility that is not usually considered (Figure 2). Table 1 summarizes all these estimates. One important conclusion of this essay is the recognition that new approaches are needed to estimate both of these fluxes.

Continental crust formation and the supercontinent cycle

Because crustal growth and destruction happen continuously if plate tectonic processes (and non-plate tectonic processes) are continuous, random collisions through Earth history should not appear as concentrated pulses of crustal growth or loss. Plate tectonic processes today produce and destroy continental crust at approximately equivalent rates, but crustal growth rates may change quasi-sinusoidally, over a supercontinent cycle. The supercontinent cycle (Nance *et al.* 1988) involves the fragmentation of supercontinents accompanying the formation of new oceans, dispersal of fragments, and recombination to form a new supercontinent. A single cycle encompasses 300–500 million years; we are approximately 200 million years (mid-way) through the supercontinent cycle that began with the break-up of Pangea approximately 200 Ma. Because the supercontinent cycle requires major tectonic reorganizations both at its beginning and end and because it greatly impacts sealevel and thus climate as well as evolution, it is sometimes called ‘The Pulse of the Planet’.

There is thus a *yin–yang* associated with crustal growth and destruction over a supercontinent cycle, with *crustal growth* dominating early and *crustal destruction* dominating late. Juvenile crustal growth should be highest during the early stages (break-up and dispersal) and crustal destruction should be greatest near the end of the cycle (Figure 7). Early rapid growth reflects increased volcanic rifted margin and hotspot crustal additions as the supercontinent ruptures and as arcs amass over new subduction zones, which must form to allow the continents to separate.

Crustal growth rate decreases over a supercontinent cycle. Early pulses of magmatic additions to the crust quickly wane, as volcanic rifted margins die and subside to form passive continental margins and as arcs mature. Crustal growth also slows because the total length of convergent margins decreases as arcs collide, reducing the cumulative length of convergent margins and thus the rate of juvenile crust production.

At the same time that crustal growth rate decreases over the supercontinent cycle, the crustal destruction rate rises. There are three processes of crustal destruction that are enhanced late in a supercontinent cycle: (1) creation and erosion of high collisional mountains, which removes upper crust and transports this detritus to the sea, where much of it is ultimately lost by subduction; (2) continental subduction, where continental crust is lost wholesale to the deep mantle; and (3) enhanced foundering, where thickening of continental crust enhances detachment of lower continental crust and descent of this material into the underlying mantle.

The conclusion that crustal growth and destruction varies systematically over a supercontinent cycle seems inescapable to us. Crustal growth should dominate early, crustal destruction should dominate late. A significant conundrum to resolve is why the rock record seems to record that crustal growth is at a maximum when supercontinents are thought to have formed, especially for postulated Precambrian supercontinents (Condie 2004; Hawkesworth *et al.* 2009). Based on our analysis of modern tectonic processes, crustal growth is greatest shortly after the supercontinent breaks up and is at a minimum as and after the supercontinent tectonically congeals. Hawkesworth *et al.* (2009) argue that the apparent abundance of igneous rocks during times of inferred supercontinent formation reflects the fact that other tracts of juvenile crust – such as arcs and volcanic rifted margins – are destroyed. We are puzzled by this posit, because the immense size of juvenile arcs – approximately 200–300 km across and up to 30 km thick – are too great to be easily obliterated. For example, during the past approximately 175 million years the long-term rate of crustal truncation (i.e. landward migration of the trench axis toward a fixed onshore reference) of northern Chile margin has averaged approximately 1.5 km/million years (275 km in 175 million years; Scholl and von Huene 2007). This rate of truncation effected by subduction erosion is quite slow, approximately 2.5% of the orthogonal underthrusting rate. At this long-term rate, removal of an accreted or obducted intra-oceanic arc massif like the Talkeenta arc of Alaska would take 150–200 million years. More inboard continental margin arcs (Andean arcs), including their forearcs, would take as long. Yet those of Mesozoic to early Tertiary age persist around much of the North Pacific rim. Furthermore, removal of forearc material can force the magmatic arc to retreat, but it will not disappear. Subduction erosion requires subduction, which engenders arc igneous activity, so the arc may retreat but will never disappear. We conclude that juvenile crust after accretion is commonly preserved. For this reason, we are perplexed by the inference from Precambrian orogens that crustal growth is maximized during times of supercontinent formation as a result of plate tectonic processes. Quite clearly, much remains to be understood about *yin–yang* processes of crustal addition loss, as well as when and how these processes have interacted to construct the carapace of continental crust that presently occupies about a third of the Earth's surface.

Conclusions

The accumulation of continental crust, presently about 7×10^9 km^3, reflects the net outcome of tectonic processes that add silica-enriched, mantle-derived igneous material to the Earth's surface while returning some of it back to the mantle. We refer to the twain of continental crust production and destruction in the terms of the ancient Chinese philosophical concept of *yin–yang*, which inextricably links processes of birthing and dying, creating and destroying.

About one-third of the Earth's surface is covered with continental crust that nurtured the evolution of all land and plant animals and the creatures of our submerged continental margins. It is widely agreed that continental crust is, in the main, a approximately 40 km-thick stack of igneous rock of average intermediate to felsic composition that evolved from mafic melts extracted from the underlying ultramafic mantle. Presently, and certainly for the Phanerozoic and probably longer, new continental crust has in particular, but not exclusively, piled up as arc massifs above intra-oceanic and continental margin subduction zones. Other but less voluminous additions of continental crust occur at volcanic rifted passive margins and hotspot injections of mafic magma into continental masses (Figure 3). Globally, during at least Phanerozoic time, the average addition of juvenile arc crust is estimated at approximately 2.7 km^3/year or 2.7 AU and at all others sites at approximately 0.5 AU for a total of 3.2 AU (Table 1). The relative proportions are thus approximately 6 to 1 (85% at subduction zones and 15% elsewhere).

Above subducting slabs, induced melting of dense mantle peridotite (~3.3 g/cm^3) generates less dense (~2.9–3.0 g/cm^3) mafic magma that ascends to build the initial rock framework of arc crust. At newly established subduction zones initial production (addition) rates are prodigious (300–500 km^3/million years/km of arc) and equal to that of slow spreading centres. With time, arc production rate wanes by a factor of 10 or more to that characteristic of a mature subduction zone (30–35 km^3/million years/km). Also with time, cycles of additions and reheatings by new batches of mafic melts generate derivative, silica-enriched crust of lower density (2.7–2.8 g/cm^3). These built the upper stories of the arc massif that are compositionally continental in character. To reach the average intermediate composition of continental crust it seems imperative that large quantities of residual, melt-produced ultramafic minerals of high density must sink beneath the arc back into the mantle. Building an arc massif with an average intermediate continental composition is a *yin–yang* process, requiring both magmatic additions from the mantle to the crust as well as the foundering loss of dense, iron-, and magnesium-rich ultramafic cumulates.

Similarly, masses of continental crust are continuously reduced by subaerial and tectonic processes. The debris of reduction is recycled to the mantle mainly at subduction zones. For the upper plate, the principal recycling mechanisms are sediment subduction and subduction erosion, and for the lower plate, the detachment loss of continental crust dragged into the mantle at crust-suturing subduction zones by subducting oceanic crust (Scholl and von Huene 2009; Figure 3). Crustal recycling also occurs away from subduction zones beneath thickened piles of continental crust where pressure-induced densification of lower crustal rocks cause them to tectonically down-well or founder into the underlying mantle lithosphere. For the Phanerozoic, and possibly longer, the long-term global loss of continental crust is estimated at approximately 3.2 AU. The bulk of which, 2.9 AU, is effected by sediment subduction (1.1 AU) and subduction erosion (1.8 AU) (Figure 3: Table 1).

A *yin–yang* balance of additions and subtractions thus appears to have been struck at approximately 3.2 AU. But the most poorly assessed components are linked to destructive losses, in particular by lower plate crustal subduction. During the Phanerozoic a net loss of crust is possible.

Once compiled, rafts of continental crust are tectonically disrupted, dispersed, and episodically reassembled into super continents, such as the coalescing of continental crust that produced Pangea about 240 Ma, only to break apart about 200 Ma to generate the Atlantic Ocean, close the Tethys Ocean along crust-suturing subduction zones separating Eurasia, Africa, and India, and the Indian Ocean separating Australia from SE Asia. Break-up of supercontinents should be attended by the creation of new subduction zones at the leading edges of dispersing fragments and also have volcanic rift margins trailing them. It would seem that the maximum of new continental crust production at magmatic arcs would occur during the break-up and dispersal phase of the supercontinent cycle.

In opposition, the maximum rates of crustal destruction would be expected to attend the coagulating of crustal rafts into a new supercontinent. We are thus puzzled by the observation that spurts of addition of juvenile magmatic crust occur during times of assembled supercontinents (Condie 2004; Hawkesworth *et al.* 2009). It seems clear that with respect to the long-term (hundreds of millions of years) balance of additions of juvenile crustal and recycling subtractions that much improved observational information is needed from the rock record of modern convergent margins and that stored in our continents. This quest is increasingly more difficult to carry out for deeper geologic time because the processes of crustal destruction increasingly dominate the preserved record of once-created continental crust.

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References

Anderson, D., 2005, Large igneous provinces, delamination, and fertile mantle: Elements, v. 1, p. 271–275.

Annen, C., Blundy, J.D., and Sparks, R.S.J., 2006, The genesis of intermediate and silicic magmas in deep crustal hot zones: Journal of Petrology, v. 47, p. 505–539.

Armstrong, R.L., 1991, The persistent myth of crustal growth: Australian Journal of Earth Sciences, v. 38, p. 613–630.

Bédard, J.H., 2006, A catalytic delamination-driven model for coupled genesis of Archaean crust and sub-continental lithospheric mantle: Geochimica et Cosmochimica Acta, v. 70, p. 1188–1214.

Behn, M.D., and Kelemen, P.D., 2006, Stability of arc lower crust: Insights from the Talkeetna arc section, south central Alaska, and the seismic structure of modern arcs: Journal of Geophysical Research, v. 111, p. B11207, doi:10.1029/2006JB004327.

Bowring, S.A., and Housh, T., 1995, The Earth's early evolution: Science, v. 269, p. 1535–1540.

Bowring, SA., and Williams, I.S., 1999, Priscoan (4.00–4.03) orthogneisses from northwestern Canada: Contributions to Mineralogy and Petrology, v. 134, p. 3–16.

Bryan, S.E., Riley, T.R., Jerram, D.A., Stephens, C.J., and Leat, P.T., 2002, Silicic volcanism, an undervalued component of large igneous provinces and volcanic rifted margins, *in* Menzies, M.A., Klemperer, S.L., Ebinger, C.J., and Baker, J., eds., Volcanic rifted margins: Boulder, CO, Geological Society of America Special Paper 362, p. 97–118.

Cagnioncle, A.-M.E., Parmentier, M., and Elkins-Tanton, L.T., 2007, Effect of solid flow above a subducting slab on water distribution and melting at convergent plate boundaries: Journal of Geophysical Research, v. 112, p. B09402, doi:10.1029/2007JB004934.

Calvert, A.J., Klemperer, S.L., Takahashi, N., and Kerr, B.C., 2008, Three-dimensional crustal structure of the Mariana island arc from seismic tomography: Journal of Geophysical Research, v. 113, p. B01406, doi:10.1029/2007JB004939.

Carr, M.J., Saginor, I., Alvarado, G.E., Bolge, L.L., Lindsay, F.N., Milidakis, K., Turrin, B.D., Feigenson, M.D., and Swisher, C.C., III, 2007, Element fluxes from the volcanic front of Nicaragua and Costa Rica: Geochemistry, Geophysics, Geosystems, v. 8, p. Q06001, doi:10.1029/2006GC001396.

Christensen, N.I., and Mooney, W.D., 1995, Seismic velocity structure and composition of the continental crust: A global view: Journal of Geophysical Research, v. 100, no. B6, p. 9761–9788.

Clendenen, W.S., Fisher, D., and Byrne, T., 2003, Cooling and exhumation history of the Kodiak accretionary prism, southwest Alaska, *in* Roeske, S.M., Pavlis, T.L., and Sisson, V.T., eds., Geology of a transpressional orogen developed during ridge-trench interaction along the North Pacific Margin: Boulder, CO, Geological Society of America Special Paper 371, p. 71–88.

Clift, P.D., Pavlis, T., DeBari, S.M., Draut, A.E., Rioux, M., and Kelemen, P.B., 2005, Subduction erosion of the Jurassic Talkeetna-Bonanza arc and the Mesozoic accretionary tectonics of western North America: Geology, v. 33, p. 881–884.

Clift, P.D., Pecher, I., Kukowski, N., and Hampel, A., 2003, Tectonic erosion of the Peruvian forearc, Lima Basin, by subduction and Nazca Ridge collision: Tectonics, v. 22, no. 3, p. 1023, doi:10.1029/2002TC001386.

Clift, P.D., Schouten, H., and Vannucchi, P., 2009, Arc-continent collsions, sediment recycling and the maintenance of the continental crust, *in* Cawood, P., and Kröner, A., eds., Accretionary orogens in space and time: Geological Society of London, Special Publication 318, p. 75–103.

Clift, P.D., and Vannucchi, P., 2004, Controls on tectonic accretion versus erosion in subduction zones: Implications for the origin and recycling of the continental crust: Reviews of Geophysics, v. 42, p. RG2001, doi:10.1029/2003RG000127.

Cloos, M., 1993, Lithospheric buoyancy and collisional orogenesis: Subduction of oceanic plateaus, continental margins, island arcs, spreading ridges, and seamounts: Geological Society of America Bulletin, v. 105, p. 715–737.

Cloos, M., Sapiie, B., van Ufford, A.Q., Weiland, R.J., Warren, P.Q., and McMahon, T.P., 2005, Collisional delamination in New Guinea: The geotectonics of subducting slab breakoff: Geological Society of America Special Paper 400, 51 p.

Cloos, M., and Shreve, R.L., 1988, Subduction-channel model of prism accretion, melange formation, sediment subduction, and subduction erosion at convergent plate margins: 1. Background and description: Pure and Applied Geophysics, v. 128, p. 455–500.

Coats, R.R., 1962, Magma type and crustal structure in the Aleutian Arc, *in* McDonald, G.A., and Kuno, H., eds., crust of the Pacific Basin: American Geophysical Union, Monograph, v. 6, p. 92–109.

Cogley, J.G., 1984, Continental margins and the extent and number of continents: Reviews of Geophysics and Space Physics, v. 22, p. 101–122.

Collot, J.W., Agudelo, W., Ribodetti, A., and Marcaillou, B., 2008, Origin of a crustal splay fault and its relation to the seismogenic zone and underplating at the erosional north Ecuador–south Colombia oceanic margin: Journal of Geophysical Research, v. 113, p. B12102, doi:10.1029/2008JB005691.

Condie, K.C., 2002, The supercontinent cycle: Are there two patterns of cyclicity? Journal of African Earth Sciences, v. 35, p. 79–183.

Condie, K.C., 2004, Supercontinents and superplume events: Distinguishing signals in the geologic record: Physics of the Earth and Planetary Interiors, v. 146, p. 319–332.

Condie, K.C., 2007, Accretionary orogens in space and time: Geological Society of America Memoir 200, p. 145–158.

Condie, K.C., and Kröner, A., 2008, When did plate tectonics begin? Evidence from the geologic record, *in* Condie, K.C., and Pease, V., eds., When did plate tectonics begin on planet Earth: Boulder, CO, Geological Society of America Special Paper 440, p. 1–14.

Condie, K.C., O'Neill, C., and Aster, R.C., 2009, Evidence and implications for a widespread magmatic shut-down for 250 My on Earth: Earth Planet Science Letters, v. 282, p. 294–298.

DeBari, S. M., and Sleep, N. H., 1991, High-Mg, low-Al bulk composition of the Talkeetna island arc, Alaska: Implications for primary magmas and the nature of arc crust: Geological Society of America Bulletin, v. 103, p. 37–47, doi: 10.1130/0016-7606.

DeCelles, P.G., Robinson, D.M., and Zandt, G., 2002, Implications of shortening in the Himalayan fold-thrust belt for uplift of the Tibetan Plateau: Tectonics, v. 21, p. 1062, doi:10.1029/2001TC001322.

Defant, M.J., and Drummond, M.S., 1990, Derivation of some arc magmas by melting of young subducted lithosphere: Nature, v. 347, p. 662–665.

Dickinson, W.R., 2009, Anatomy and global context of the North American Cordillera, *in* Kay, S.M., Ramos, V.A., and Dickinson, W.R., eds., Backbone of the Americas: Shallow subduction, plateau uplift, and ridge and terrane collision: Geological Society of America Memoir 204, p. 1–29.

Dickinson, W.R., and Sealy, D.R., 1979, Structure and stratigraphy of forearc regions: AAPG Bulletin, v. 63, p. 2–31.

Dimalanta, C., Taira, A., Yumul, G.P., Jr., Tokuyama, H., and Mochizuki, K., 2002, New rates of western Pacific island arc magmatism from seismic and gravity data: Earth and Planetary Science Letters, v. 202, p. 105–115.

Ducea, M.N., 2002, Constraints on the bulk composition and root foundering rates of continental arcs: A California arc perspective: Journal of Geophysical Research, v. 107, no. B11, p. 2304, doi:10.1029/2001JB000643.

Elkins-Tanton, L.T., 2007, Continental magmatism, volatile recycling, and a heterogeneous mantle caused by lithospheric gravitational instabilities: Journal of Geophysical Research, v. 112, p. B03405, doi:10.1029/2005JB004072.

Ernst, W.G., 2006, Preservation/exhumation of ultrahigh-pressure subduction complexes: Lithos, v. 92, p. 321–335.

Ernst, W.G., 2007, Petrotectonics of ultrahigh-pressure crustal and upper-mantle rocks – Implications for Phanerozoic collisional orogens, *in* Sears, J.W., Harms, T.A., and Evenchick, C.A., eds., Whence the mountains? Inquiries into the evolution of orogenic systems: A volume in honor of Raymond A. Price: Boulder, CO, Geological Society of America Special Paper 433, p. 27–49.

Ernst, W.G., 2009, Archean plate tectonics, rise of Proterozoic supercontinentality and onset of regional, episodic stagnant-lid behavior: Gondwana Research, v. 15, p. 243–253.

Fliedner, M.M., and Klemperer, S.L., 2000, Crustal structure transition from oceanic arc to continental arc, eastern Aleutian Islands and Alaska Peninsula: Earth and Planetary Science Letters, v. 179, p. 567–579.

Gao, S., Jin, Z., Kern, H., Luo, T., and Zhao, Z., 1998, How mafic is the lower continental crust?: Earth and Planetary Science Letters, v. 161, no. 1–4, p. 101–117.

Garrido, C.J., Bodinier, J.-L., Dhuime, B., Bosch, D., Chanefo, I., Bruguier, O., Hussain, S.S., Dawood, H., and Burg, J.-P., 2007, Origin of the island arc Moho transition zone via melt-rock reaction and its implications for intracrustal differentiation of island arcs: Evidence from the Jijal complex (Kohistan complex, northern Pakistan): Geology, v. 35, p. 683–686.

Glazner, A.F., 1991, Plutonism, oblique subduction, and continental growth: An example from the Mesozoic of California: Geology, v. 19, p. 784–786.

Gögüs, O., and Pysklywec, R.N., 2008, Near-surface diagnostics of dripping or delaminating lithosphere: Journal of Geophysical Research, v. 113, p. B11404, doi:10.1029/2007JB005123.

Greene, R., DeBari, S.M., Kelemen, P.B., Blusztajn, J., and Clift, P.D., 2006, A detailed geochemical study of Island Arc Crust: The Talkeetna Arc Section, South–Central Alaska: Journal of Petrology, v. 47, p. 1051–1093.

Grove, M., Bebout, G.E., Jacobson, C.E., Barth, A.P., Kimbrough, D.L., King, R.L., Zou, H., Lovera, O.M., and Gehrels, G.E., 2008. The Catalina Schist: Evidence for middle Cretaceous subduction erosion of southwestern North America, *in* Draut, A.E., Clift, P.D., and Scholl, D.W., Formation and applications of the sedimentary record in arc collision zones: Boulder, CO, Geological Society of America Special Paper 436, p. 335–361.

Grove, T.L., Till, C.B., Lev, E., Chatterjee, N., and Médard, E., 2009, Kinematic variables and water transport control the formation and location of arc volcanoes: Nature, v. 459, p. 694–697.

Hacker, B.R., 2008, H_2O subduction beyond arcs: Geochemistry, Geophysics, Geosystems, v. 9, p. Q03001, doi:10.1029/2007GC001707.

Hall, C., Gurnis, M., Sdrolias, M., Lavier, L.L., and Müller, R.D., 2003, Catastrophic initiation of subduction following forced convergence across fracture zones: Earth and Planetary Science Letters, v. 212, p. 15–30.

Harrison, T.M., 2009, The Hadean Crust: Evidence from >4 Ga Zircons: Annual Reviews of Earth and Planetary Sciences, v. 37, p. 479–505.

Harrison, T.M., Copeland, P., Kidd, W.S.F., and Yin, A., 1992, Raising Tibet: Science, v. 255, p. 1663–1670.

Haschke, M., and Günther, A., 2003, Balancing crustal thickening in arcs by tectonic vs. magmatic means: Geology, v. 31, p. 933–936.

Hawkesworth, C.J., Cawood, P., Kemp, T., Storey, C., and Dhuime, B., 2009, Geochemistry: A matter of preservation: Science, v. 323, p. 49–50.

Hawkesworth, C.J., and Kemp, A.I.S., 2006, The differentiation and rates of generation of the continental crust: Chemical Geology, v. 226, p. 134–143.

Herzburg, C.T., Fyfe, W.S., and Carr, M.J., 1983, Density constraints on the formation of the continental Moho and crust: Contributions to Mineralogy and Petrology, v. 84, p. 1–5.

Hildebrand, R.S., and Bowring, S.A., 1999, Crustal recycling by slab failure: Geology, v. 27, p. 11–14

Hildreth, W., 2007, Quaternary magmatism in the cascades – geologic perspectives: U.S. Geological Survey Professional Paper 1744, 125 p.

Holbrook, W.S., Lizarralde, D., McGeary, S., Bangs, N., and Diebold, J., 1999, Structure and composition of the Aleutian Island arc and implications for continental crustal growth: Geology, v. 27, p. 31–34.

Jackson, J., 2002, Strength of the continental lithosphere: Time to abandon the jelly sandwich? GSA Today, v. 12, p. 4–10.

Jacobsen, S.B., 1988, Isotopic constraints on crustal growth and recycling: Earth and Planetary Science Letters, v. 90, p. 315–329.

Jicha, B., Scholl, D.W., Singer, B.S., Yogodzinski, G.M., and Kay, S.M., 2006, Revised age of Aleutian Island Arc formation implies high rate of magma production: Geology, v. 34, p. 661–664.

Johnson, M.R.W., 2002, Shortening budgets and the role of continental subduction during the India-Asia collision: Earth-Science Reviews, v. 59, p. 101–123.

Johnson, E.R., Wallace, P.J., Cashman, K.V., Granados, H.D., and Kent, A.J.R., 2008, Magmatic volatile contents and degassing-induced crystallization at Volcán Jorullo, Mexico: Implications for melt evolution and the plumbing system of monogenetic volcanoes: Earth and Planetary Science Letters, v. 269, p. 478–487.

Jull, M., and Kelemen, P.B., 2001, On the conditions for lower crustal convective instability: Journal of Geophysical Research, v. 106, no. B, p. 6423–6446.

Kaneko, Y., Katayama, I., Yamamoto, H., Misawa, K., Ishikawa, M., Rehman, H.U., Kausar, A.B., and Shirashi, K., 2003, Timing of Himalayan ultrahigh-pressure metamorphism: Sinking rate and subduction angle of the Indian continental crust beneath Asia: Journal of Metamorphic Geology, v. 21, p. 589–599.

Kay, R.W., and Kay, S.M., 2008, The Armstrong Unit (AU = km^3/Yr) and processes of crust-mantle mass flux: Geochimica et Cosmochimica Acta, v. 72, p. A455 (abstract).

Kelemen, P.B., 1995, Genesis of high Mg# andesites and the continental crust: Contributions to Mineralogy and Petrology, v. 120, p. 1–19.

Kelemen, P.B., Hanghøj, K., and Greene, A.R., 2003, One view of the geochemistry of subduction-related magmatic arcs with an emphasis on primitive andesite and lower crust, *in* Rudnick, R.L., ed., The crust. Treatise on geochemistry: Oxford, Elsevier-Pergamon, p. 593–659.

Kimura, J.-I., and Stern, R.J., 2008, Neogene volcanism of the Japan Island Arc: The K-h relationship revisited, *in* Circum-Pacific Tectonics, Geologic Evolution, and Ore Deposits Symposium-Proceedings Volume, Arizona Geological Society Digest, p. 187–202.

Kimura, J.-I., and Yoshida, T., 2006, Contributions of slab fluid, mantle wedge and crust to the origin of Quaternary lavas in the NE Japan arc: Journal of Petrology, v. 47, p. 2185–2232.

Kodaira, S., Sato, T., Takahashi, N., Miura, S., Tamura, Y., Tatsumi, Y., and Kaneda, Y., 2007, New seismological constraints on growth of continental crust in the Izu-Bonin intra-oceanic arc: Geology, v. 35, p. 1031–1034.

Kukowski, N., and Oncken, O., 2006, Subduction erosion – the 'Normal' mode of Fore-Arc material transfer along the Chilean Margin? The Andes, Frontiers in Earth Science: Berlin, Springer-Verlag, p. 217–236.

Lapierre, H., Bosch, D., Tardy, M., and Strulk, L.C., 2003, Late Paleozoic and Triassic plume-related magmas in the Canadian Cordillera played a key role in continental crust growth: Chemical Geology, v. 201, p. 5–89.

Leech, M.L., Singh, S., Jain, A.K., Klemperer, S.L., and Manickavasagam, R.M., 2005, The onset of India – Asia continental collision: Early, steep subduction required by the timing of UHP metamorphism in the western Himalaya: Earth and Planetary Science Letters, v. 234, p. 83–97.

Liou, J.G., Zhang, R.V., and Ernst, W.G., 2007, Very high-pressure orogenic garnet peridotites: Proceedings of the National Academy of Sciences, v. 104, p. 9116–9121.

Mann, P., and Taira, A., 2004, Global tectonic significance of the Solomon Islands and Ontong Java Plateau convergent zone: Tectonophysics, v. 389, p. 137–190.

McCurry, M., Hayden, K.P., Morse, L.H., and Mertzmann, S., 2008, Genesis of post-hotspot, A-type rhyolite of the Eastern Snake River Plain volcanic field by extreme fractional crystallization of olivine tholeiite: Bulletin of Volcanology, v. 70, p. 361–383.

McQuarrie, N., Stock, J.M., Verdel, C., and Wernicke, B.P., 2003, Cenozoic evolution of Neotethys and implications for the causes of plate motions: Geophysical Research Letters, v. 30, p. 2036, doi:10.1029/2003GL017992.

Menzies, M.A., Klemperer, S.L., Ebinger, C.J., and Baker, J., 2002, Characteristics of volcanic rifted margins. Volcanic Rifted Margins: Boulder, CO, Geological Society of America Special Paper 362, p. 1–14.

Mjelde, R., Raum, T., Murai, Y., and Takanami, T., 2007, Continent-ocean-transitions: Review, and a new tectonomagmatic model of the Vøring Plateau, NE Atlantic: Journal of Geodynamics, v. 43, p. 374–392.

Moore, J.C., Diebold, J., Fisher, M.A., Sample, J., Brocher, T., Talwani, M., Ewing, J., von Huene, R., Rowe, C., Stone, D., Stevens, C., and Sawyer, D., 1991, EDGE deep seismic reflection transect of the eastern Aleutian arc-trench layered lower crust reveals underplating and continental growth: Geology, v. 19, p. 420–424.

Müller, R.D., Sdrolias, M., Gaina, C., and Roest, W.R., 2008, Age, spreading rates, and spreading asymmetry of the world's ocean crust: Geochemistry, Geophysics, Geosystems, v. 9, p. Q04006, doi:10.1029/2007GC001743.

Müntener, O., and Ulmer, P., 2006, Experimentally derived high-pressure cumulates from hydrous arc magmas and consequences for the seismic velocity structure of lower arc crust: Geophysical Research Letters, v. 33, p. L21308, doi:10.1029/2006GL027629.

Nance, R.D., Worsley, T.R., and Moody, J.B., 1988, The supercontinent cycle: Scientific American, v. 259, no. 1, p. 72–79.

O'Brien, P.J., Zotov, N., Law, R., Khan, M., Ahmed, M., and Qasim, J.M., 2001, Coesite in Himalayan eclogite and implications for models of India-Asia collision: Geology, v. 29, p. 435–438.

O'Neill, C., Lenardic, A., Moresi, L., Torsvik, T.H., and Lee, C.-T.A., 2007, Episodic Precambrian subduction: Earth and Planetary Science Letters, v. 262, p. 552–562.

Plank, T., 2005, Constraints from Thorium/Lanthanum on sediment recycling at subduction zones and the evolution of the continents: Journal of Petrology, v. 46, p. 921–944.

Portnyagin, M., Savelyev, D., Hoernle, K., Hauff, F., and Garbe-Schönberg, D., 2008, Mid-Cretaceous Hawaiian tholeiites preserved in Kamchatka: Geology, v. 36, p. 903–906.

Redfield, T.F., Scholl, D.W., and Fitzgerald, P.G., 2007, Escape tectonics and the extrusion of Alaska: Past, present, and future: Geology, v. 35, p. 1039–1042.

Reilinger, R., McClusky, S., Vernant, P., and 21 others, 2006, GPS constraints on continental deformation in the Africa-Arabia-Eurasia continental collision zone and implications for the dynamics of plate interactions: Journal of Geophysical Research, v. 111, p. B05411, doi:10.1029/2005JB004051.

Reymer, A., and Schubert, G., 1984, Phanerozoic addition rates to the continental crustal growth: Tectonics, v. 3, p. 63–77.

Richards, J.P., and Kerrich, R., 2007, Adakite-like rocks: Their diverse origins and questionable role in metallogenesis: Economic Geology, v. 102, p. 537–576.

Rioux, M., Hacker, B., Mattinson, J., Kelemen, P., Blusztajn, J., and Gehrels, G., 2007, Magmatic development of an intra-oceanic arc: High-precision U-Pb zircon and whole-rock isotopic analyses from the accreted Talkeetna arc, south-central Alaska: Geological Society of America Bulletin, v. 119, p. 1168–1184.

Rudnick, R.L., and Fountain, D., 1995, Nature and composition of the continental crust: A lower crustal perspective: Reviews of Geophysics, v. 33, p. 267–309.

Rudnick, R.L., and Gao, S., 2003, Composition of the continental crust, *in* Treatise on geochemistry: Oxford, Elsevier-Pergamon, v. 3, p. 1–64.
Saleeby, J., 2003, Segmentation of the Laramide Slab – Evidence from the southern Sierra Nevada region: Geological Society of America Bulletin, v. 115, p. 655–668.
Saleeby, J., Ducea, M., and Clemens-Knott, D., 2003, Production and loss of high-density batholithic root, southern Sierra Nevada, California: Tectonics, v. 22, no. 6, p. 1064, doi:10.1029/2002TC001374.
Sample, J.C., and Karig, D.E., 1982, A volcanic production rate for the Mariana island arc: Journal of Volcanology and Geothermal Research, v. 13, p. 73–82.
Sample, J.C., and Moore, J.C., 1987, Structural style and kinematics of an underplated slate belt, Kodiak and adjacent islands, Alaska: Geological Society of America Bulletin, v. 99, p. 7–20.
Scambelluri, M., Pettke, T., and van Roermund, H.L.M., 2008, Majoritic garnets monitor deep subduction fluid flow and mantle dynamics: Geology, v. 36, p. 59–62.
Scholl, D.W., 2007, Viewing the tectonic evolution of the Kamchatka-Aleutian (KAT) connection with an Alaska crustal extrusion perspective, *in* Eichelberger, J.C., Gordeev, E., Izbekov, P., Kasahara, M., and Lees J., eds., Volcanism and subduction the Kamchatka region: American Geophysical Union, Monograph 172, p. 3–35.
Scholl, D.W., and von Huene, R., 2009, Implications of estimated magmatic additions and recycling losses at the subduction zones of accretionary (non-collisional) and collisional (suturing) orogens, *in* Cawood, P., and Kröner, A., eds., Accretionary orogens in space and time: Geological Society of London, Special Publication 318, p. 105–125.
Scholl, D., and von Huene, R., 2007, Crustal recycling at modern subduction zones applied to the past – Issues of growth and preservation of continental basement crust, mantle geochemistry, and supercontinent reconstruction, *in* Hatcher, R.D., Jr., Carlson, M.P., McBride, J.H., and Martínez Catalán, J.R., eds., 4-D framework of continental crust: Geological Society of America Memoir 200, p. 9–32.
Scholl, D.W., von Huene, R., Vallier, T.L., and Howell, D.G., 1980, Sedimentary masses and concepts about tectonic processes at underthrust ocean margins: Geology, v. 8, p. 564–568.
Shillington, D.J., Van Avendonk, H.J.A., Holbrook, W.S., Kelemen, P.B., and Hornbach, M.J., 2004, Composition and structure of the central Aleutian island arc from arc-parallel wide-angle seismic data: Geochemistry, Geophysics, Geosystems, v. 5, p. Q10006, doi:10.1029/2004GC000715.
Smrekar, S.E., and Stofan, E.R., 1997, Corona formation and heat loss on Venus by coupled upwelling and delamination: Science, v. 277, p. 1289–1294.
Stern, R.J., 2002, Subduction zones: Reviews of Geophysics, v. 40, DOI: 10.1029/2001RG000108.
Stern, R.J., 2004, Subduction initiation: Spontaneous and induced: Earth and Planetary Science Letters, v. 226, p. 275–292.
Stern, R.J., 2008, Modern-style plate tectonics began in neoproterozoic time: An alternative interpretation of Earth's tectonic history, *in* Condie, K., and Pease, V., eds., When did plate tectonics begin? Boulder, CO, Geological Society of America Special Paper 440, p. 265–280.
Stern, R.J., and Bloomer, S.H., 1992, Subduction zone infancy: Examples from the Eocene Izu-Bonin-Mariana and Jurassic California Arcs: Geological Society of America Bulletin, v. 104, p. 1621–1636.
Suyehiro, K., Takahashi, N., Ariie, Y., Yokoi, Y., Hino, R., Shinohara, M., Kanazawa, T., Hirata, N., Tokuyama, H., and Taira, A., 1996, Continental crust, crustal underplating, and low-Q upper mantle beneath an Oceanic Island Arc: Science, v. 272, p. 390–392.
Takahashi, N., Kodaira, S., Klemperer, S.L., Tatsumi, Y., Kaneda, Y.S., and Suyehiro, K., 2007, Crustal structure and evolution of the Mariana intra-oceanic island arc: Geology, v. 35, p. 203–206.
Takahashi, N., Kodaira, S., Tatsumi, Y., Kaneda, Y., and Suyehiro, K., 2008, Structure and growth of the Izu-Bonin-Mariana arc crust: 1. Seismic constraint on crust and mantle structure of the Mariana arc–back-arc system: Journal of Geophysical Research, v. 113, no. B01104, doi:10.1029/2007JB005120.
Tatsumi, Y., 2005, The subduction factory: How it operates in the evolving Earth: GSA Today, v. 15, no. 7, p. 4–10.
Tatsumi, Y., Shukuno, H., Tani, K., Takahashi, N., Kodaira, S., and Kogiso, T., 2008, Structure and growth of the Izu-Bonin-Mariana arc crust: 2. Role of crust-mantle transformation and the transparent Moho in arc crust evolution: Journal of Geophysical Research, v. 113, p. B02203, doi:10.1029/2007JB005121.

Van der Voo, R., Spakman, W., and Bijwaard, H., 1999, Tethyan subducted slabs under India: Earth and Planetary Science Letters, v. 171, p. 7–20.

Vernant, P., Nilforoushan, F., Hatzfeld, D., Abbassi, M.R., Vigny, C., and Masson, F., 2004, Present-day crustal deformation and plate kinematics in the Middle East constrained by GPS measurements in Iran and northern Oman: Geophysical Journal International, v. 157, p. 381–398.

von Huene, R., and Ranero, C.R., 2009, Neogene collision and deformation of convergent margins along the Americas backbone, *in* Kay, S., Ramos, V., and Dickinson, W., eds., Backbone of the Americas: Shallow subduction, plateau uplift, and ridge collision: Geological Society of America Memoir 204, p. 67–84.

von Huene, R., Ranero, C.R., and Vannucchi, P., 2004, Generic model of subduction erosion: Geology, v. 32, p. 913–916.

von Huene, R., and Scholl, D.W., 1991, Observations at convergent margins concerning sediment subduction, subduction erosion, and the growth of continental crust: Reviews of Geophysics, v. 29, p. 279–316.

Wells, R.E., Blakely, R.J., Sugiyama, Y., Scholl, D.W., and Dinterman, P.A., 2003, Basin-centered asperities in great subduction zone earthquakes – a link between sip, subsidence and subduction erosion? Journal of Geophysical Research, v. 108, no. B10, p. 2507, doi:10.1029/2002JBOO2072.

White, S.M., Crisp, J.A., and Spera, F.J., 2006, Long-term volumetric eruption rates and magma budgets: Geochemistry, Geophysics, Geosystems, v. 7, p. Q03010, doi:10.1029/2005GC001002.

Zandt, G., Gilbert, H., Owens, T.J., Ducea, M., Saleeby, J., and Jones, C.H., 2004, Active foundering of a continental arc root beneath the southern Sierra Nevada in California: Nature, v. 431, p. 41–46.