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Continents as lithological icebergs: the importance of buoyant lithospheric roots

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Abstract

An understanding of the formation of new continental crust provides an important guide to locating the oldest terrestrial rocks and minerals. We evaluated the crustal thicknesses of the thinnest stable continental crust and of an unsubductable oceanic plateau and used the resulting data to estimate the amount of mantle melting which produces permanent continental crust. The lithospheric mantle is sufficiently depleted to produce permanent buoyancy (i.e., the crust is unsubductable) at crustal thicknesses greater than 25–27 km. These unsubductable oceanic plateaus and hotspot island chains are important sources of new continental crust. The newest continental crust (e.g., the Ontong Java plateau) has a basaltic composition, not a granitic one. The observed structure and geochemistry of continents are the result of convergent margin magmatism and metamorphism which modify the nascent basaltic crust into a lowermost basaltic layer overlain by a more silicic upper crust. The definition of a continent should imply only that the lithosphere is unsubductable over ≥ 0.25 Ga time periods. Therefore, the search for the oldest crustal rocks should include rocks from lower to mid-crustal levels.

Keywords: continental crust; basalts; subduction; lithosphere; mantle

1. Introduction

The evidence for the most ancient continental crust is derived from dating of zircons [1], which form primarily in granitic rocks. At the present time the oldest known terrestrial rocks are ~ 4.0 Ga [2] and the oldest known terrestrial minerals are ~ 4.2 Ga zircons [1]. The Earth's moon has no atmosphere and a similar bombardment history to the Earth. On the Moon, ~ 4.4 Ga basaltic crust still exists [3]. Although the surface of the Earth is much more tectonically active than the Moon, exhumed metamorphic rocks have provided the oldest terrestrial ages [2]. Zircons in buried rocks can retain their isotopic ages for billions of years under the typical P–T conditions within cratonic crust. Therefore, it is surprising that no one has found terrestrial rocks dating from before ~ 4.0 Ga. It has also been assumed that the oldest and first formed continental crust must have been sialic. If, however, the earliest crust was basaltic, it would have contained far fewer zircons. Consequently, it is important to understand whether the long term preservation of crustal rocks really requires a granitic crust.

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Early geological mappers based their estimates of average continental crustal composition on the rocks that they observed, which were largely silica and aluminum rich (e.g., granite or sial). In contrast, most rocks dredged from the ocean basins are silica and magnesium rich (e.g., basalt or sima) [3]. Sialic crust is less dense than simatic crust, providing a seeming explanation for the permanency of continental crust [4] compared to oceanic crust. However, this is an oversimplification, similar to the idea that an iceberg is kept up by the snow on its upper surface. The snow on an iceberg is less dense than the surrounding water, but the snow layer does not make the iceberg float. Rather, the iceberg floats because it has a root of ice that is less dense than the surrounding water. Continents also have a less dense surface layer (i.e., the crust) but we show that it is their buoyant mantle roots that make it possible for continents to resist subduction and to remain at the surface of the Earth. We show that granitic crust is not required for continental preservation and that the first formed continental crust was more probably basaltic in composition.

2. The composition of continents

Although the average composition of continental crust is dioritic [4,5], the crust contains large amounts of basalt. Matching of seismic velocities to lithology supports a basaltic (mafic granulite) composition for the lowest 10 km of crust, a basalt–granite (amphibolite plus granite) mixture for the middle 20 km, and a granitic composition for the upper 10 km [5]. Because igneous rocks of intermediate composition are rarely primary melts, we interpret the seismic data from the middle crust as indicating that it is composed of a mixture of granitic plus basaltic rocks. This suggests that the composite thickness of the basaltic component may be 20 km.

Most estimates of the seismic velocity of basaltic rocks have been derived from mid-ocean ridge basalts (MORBs) (see references in [5]), but hotspot and plume basalts are enriched in iron compared to MORB. For a given mafic mineral, the iron-rich end-member has a lower seismic velocity [6]. Traditionally, lower seismic velocities have been interpreted as the result of a higher silica content. However, if the iron content of the crust has been underestimated, then its silica content (and radiogenic heat production) have been overestimated. Therefore, the continental crust may contain a higher proportion of basaltic component than previously thought.

We know that continents grow laterally by the incorporation of oceanic crust formed within the ocean basins [7–9]. However, the mechanisms that turn some sections of oceanic crust into new continents are still debated. Continents may grow by accretion of oceanic plateaus, hotspot tracks, or island arcs. In the latter case, the upper crust is fairly silicic (basaltic andesites), and the lower crust is basaltic [10]. In the former cases, new continental crust is over 99% basaltic. If the major proportion of newly formed continental crust is derived from accreted oceanic plateaus and hotspot tracks, we can also assume that the continental crust is proportionately more basaltic than if it formed dominantly by accretion of island arcs.

3. Subduction character and crustal thickness

Previously, we showed that oceanic crustal thickness is directly correlated with the way a plate subducts [11]. Plates with thinner oceanic crust subduct more steeply, while those with thicker oceanic crust subduct more shallowly. Examples of these types of subduction behavior are shown in Fig. 1, a map of the Circum-Pacific. The subduction of thinner oceanic crust produces deep trenches and deep earthquakes. Subduction of thicker oceanic crust produces shallow trenches and shallow earthquakes. Because shallow subduction reduces the size of the asthenospheric wedge beneath the arc, areas of shallow subduction have either no arc volcanism or arc volcanism of an unusual composition (Fig. 2).

The observed differences in subduction style are caused by differences in the relative buoyancy of plates. Plates of the same age with thinner oceanic crust are less buoyant than plates with thicker oceanic crust. However, subducting oceanic plates with thicker crust have thicker layers of eclogite, which is denser than the asthenospheric mantle. Therefore, differences in subduction style are not directly controlled by the thickness and density of the subducting oceanic crust, but indirectly by differences in the relative buoyancy of the oceanic lithospheric mantle resulting from crustal extraction during mantle melting.

4. Lithospheric buoyancy and crustal extraction

The buoyancy of the oceanic lithosphere is largely a function of its density distribution. The density distribution is controlled primarily by the relative composition and thickness of the crust and mantle layers. Crustal thickness and mantle composition are, in turn, controlled by the mantle temperature at the location where the crust formed. Although some oceanic crust forms off-ridge, most oceanic crust forms at a spreading ridge. Hotter mantle melts more and produces thicker oceanic crust and more depleted lithospheric mantle [12,13]. Colder mantle melts less and produces thinner oceanic crust and less depleted lithospheric mantle.

When oceanic plates begin to subduct, their buoyancy can be gauged in part by their crustal thickness. Plates with thick oceanic crust are buoyant relative to the mantle for two reasons: (1) the crustal layer is thicker, and oceanic crust is less dense than mantle olivine; and (2) there is less garnet and less iron in the sub-oceanic lithospheric mantle. Garnet is the most dense of the common upper mantle minerals.

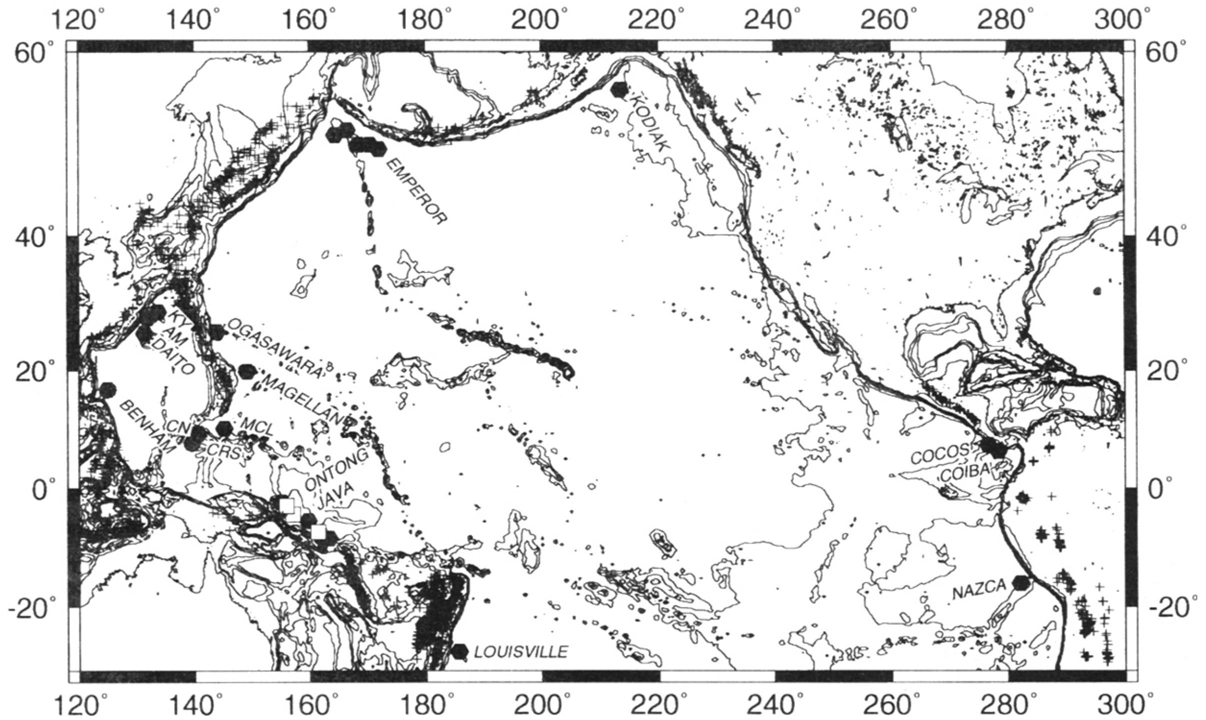


Fig. 1. Map of the Pacific Ocean basin showing the location of local topographic highs with flat subduction (hexagons) and unsubductable oceanic lithosphere (open squares). The bathymetry is contoured at 1000 m intervals. Most of the Ontong Java plateau is unsubductable. All other features have flat subduction: Kodiak seamount chain, Cocos ridge, Coiba Ridge, Nazca Ridge, Louisville Ridge, CRS (Caroline Ridge South), CN (Caroline Ridge North), MCL (McLaughlin Bank), Magellan Seamount Chain, Hawaiian–Emperor Seamount Chain, Ogasawara Plateau, Benham Rise, AM (Amami) Plateau, Oki-Daito Ridge, and KY (Palau–Kyushu) Ridge. + = Distribution of deep earthquakes (pP depth > 150 km). Areas of shallow subduction have shallower trenches and constitute local topographic highs. The majority of shallow subduction produces a broad, clearly visible zone that lacks deep earthquakes; for example, Nazca Ridge, Cocos Ridge, Coiba Ridge, Kodiak seamounts, Benham Rise, Caroline Ridge North, Caroline Ridge South, McLaughlin Bank, and the Emperor seamounts. In other areas, either where the subducting plate is especially old (Ogasawara Plateau and Magellan seamounts), and/or the region of thickened crust is narrower or more diffuse (Louisville Ridge, Palau–Kyushu and Amami), the shallowing of the Benioff zone is more local and it is not visible on this map.

Garnet also melts at a lower temperature than olivine. Partial melting of the mantle also removes iron from olivine. Mantle which has had more partial melt removed has less residual garnet and less iron and is less dense than more fertile mantle [14]; thus, it becomes more buoyant.

When subducting basalt inverts to dense eclogite, its buoyancy with respect to the asthenosphere is lost. Consequently, the buoyancy resulting from less garnet and less iron in the lithospheric mantle is the only buoyancy that can survive the onset of subduction. As a result, a subducting plate with thick oceanic crust produces a nearly flat Benioff zone because the positive buoyancy from the depleted lithospheric mantle overwhelms the negative effect of a greater thickness of dense eclogite. If the oceanic crust is over 25 km thick, the oceanic lithosphere can become too light to subduct.

The temperature gradient within a subducting plate also affects buoyancy, through the effects of thermal expansion and contraction. As a plate ages it cools. Young, hot plates (less than ~ 5 m.y. old) are buoyant regardless of their crustal thickness. However, older, colder plates are buoyant only if the crust is thicker than the present day normal thickness of ~ 7 km [15]. If the lithospheric mantle has lost enough garnet and iron, the plate will not subduct even when it is old and cold. These unsubductable oceanic plates become new continental plates.

5. Lithospheric buoyancy and subduction style

Subduction style depends primarily upon the buoyancy of the oceanic lithosphere. Previously, we described the subduction behavior of oceanic litho-

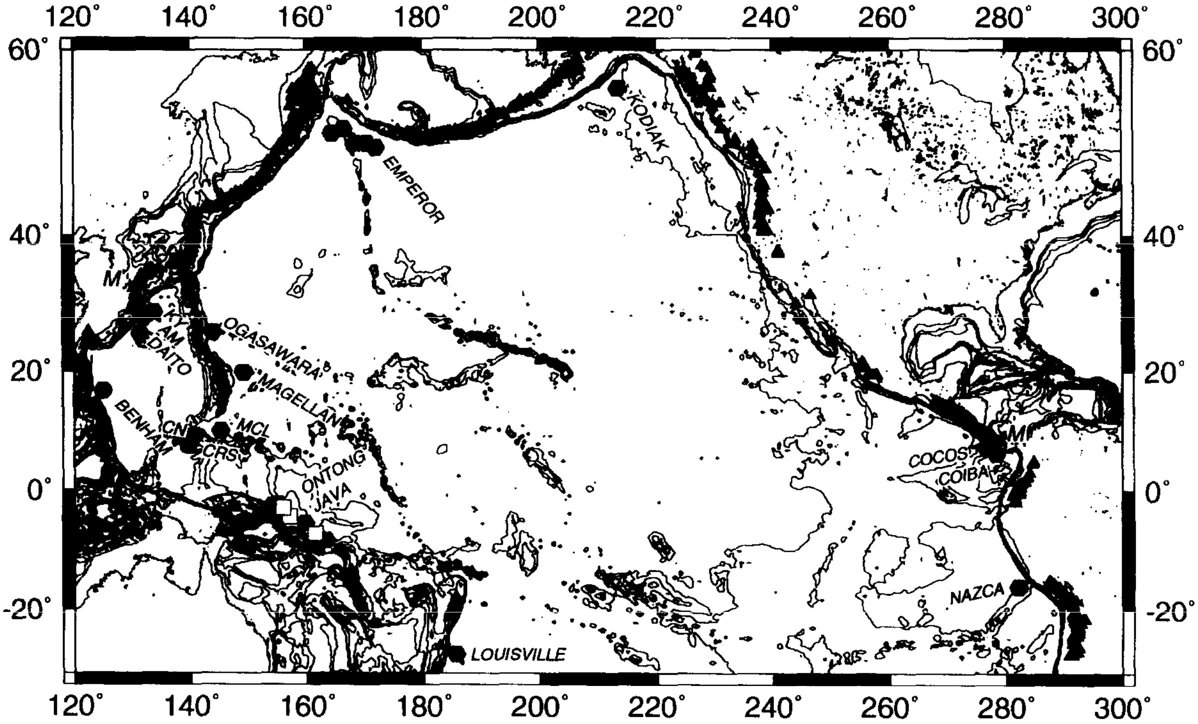


Fig. 2. Distribution of active volcanoes (filled triangles) from [43]. In areas of shallow subduction, there is either no arc volcanism or arc volcanism of a different composition (M). The unusual composition of the silicic arc volcanics in the latter areas is the result of partial melting of the basalt layer on the subducting plate. Slab melt components within arc volcanics are produced by the subduction of the Cocos and Coiba ridges [44,45] and the subduction of the Palau–Kyushu ridge [46]. Many features with shallow subduction not resolved in Fig. 1 produce clear gaps in the line of active volcanoes.

sphere which returns to the mantle. Crustal thickness is the major control on subduction style [11], with older plates (> 10–20 m.y.) subducting in two ways. Thinner oceanic crust (< 10 km thick) subducts steeply, producing steep Benioff zones, deep earthquakes, and large negative gravity anomalies [16]. Thicker oceanic crust (> 10 km thick) subducts shallowly, producing nearly flat Benioff zones, shallow earthquakes, and slightly negative gravity anomalies [17,18]. In this paper we introduce a third subduction style, in which the oceanic plate is so buoyant that it cannot return to the mantle. We then seek to determine the crustal thickness that divides oceanic plates which subduct shallowly from oceanic plates which cannot subduct.

6. Subduction styles in the Circum-Pacific region

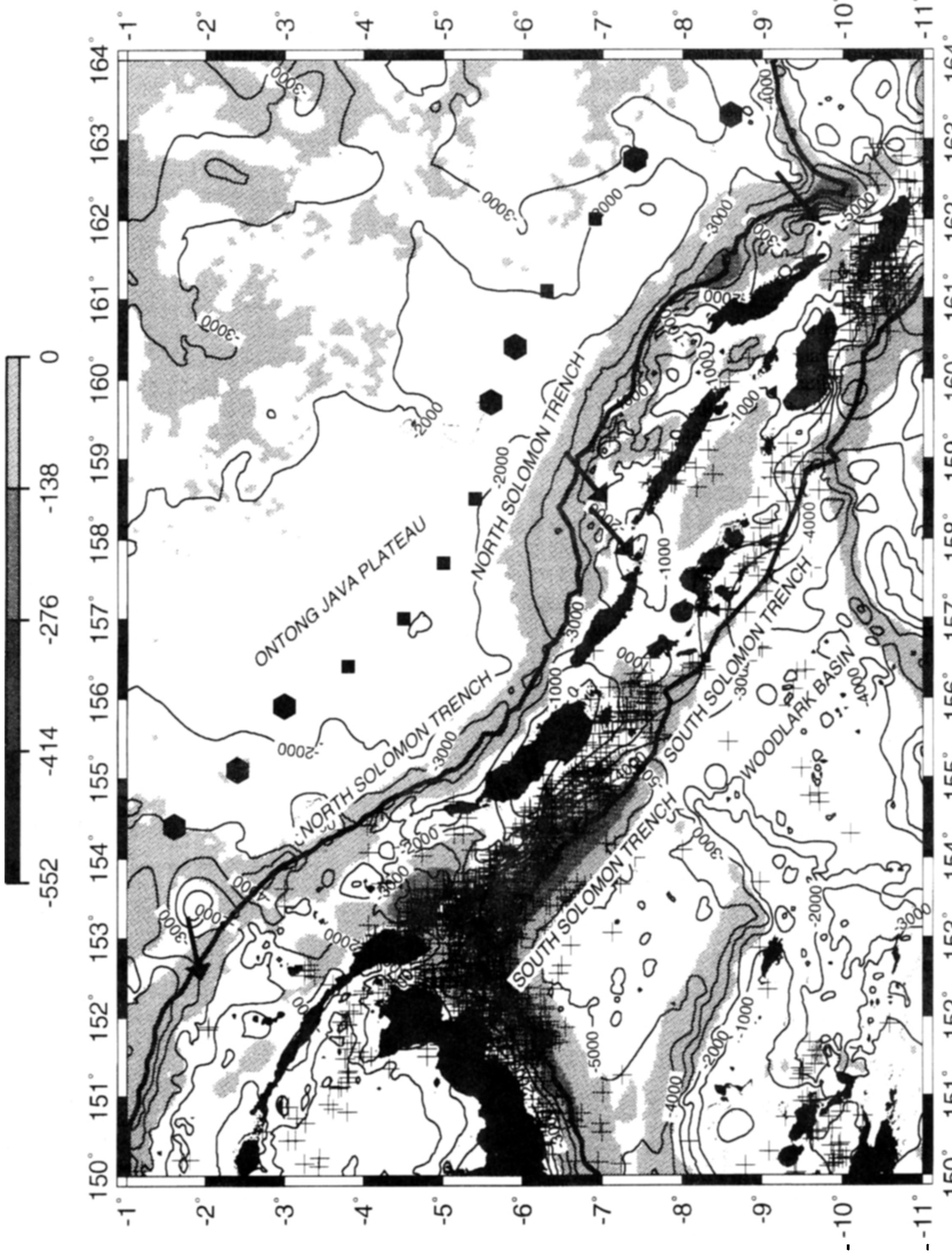
Differences in relative buoyancy of oceanic plates are reflected in seafloor depths. Areas of thicker crust riding on more buoyant plates form local topographic highs in the ocean basins (Fig. 1). Differences in the relative buoyancy of oceanic plates are also reflected in the topography and axial gravity at trenches. At a trench, some proportion of the negative gravity anomaly is produced by bending of the subducting plate beneath the overriding plate. This negative gravity anomaly will occur even when the plate is too light to subduct. For example, simple bending of the plate produces the negative gravity anomalies in the flexural moat surrounding the Hawaiian hotspot, the strongest hotspot. The most negative gravity value in the Hawaiian flexural moat is −138 mgal. If the trench axis gravity is more negative than −138 mgal, it is likely that additional forces are increasing the bending moment of the subducting slab. These additional bending forces are produced by the negative buoyancy of the subducting plate within the Benioff zone.

A plate that is too light to subduct will also stop convergence, and its shallow levels will cease to generate subduction zone earthquakes. Because the attempted subduction of the Ontong Java plateau caused the trench to reverse its polarity [19], we infer that parts of the Ontong Java plateau are unsubductable. Although there is some continued convergence and weak seismicity [20], most of the North Solomon trench axis is seismically inactive and has an axial gravity value exceeding −138 mgal (Fig. 3). We infer that the seismically inactive portions of the North Solomon trench with axial gravity exceeding −138 mgal represent areas of unsubductable oceanic lithosphere. The remainder of the Ontong Java trench contains slowly converging, nearly neutrally buoyant oceanic lithosphere that produces weak seismicity, shallowly dipping subduction, and subdued trench axis gravity anomalies. Therefore, we infer that the North Solomon trench near the Ontong Java Plateau contains two types of oceanic lithosphere: oceanic lithosphere which is subducting shallowly, and that which is unsubductable.

As the buoyancy of an oceanic plate increases, the relative difference in the topographic elevation between a normal plate and a more buoyant plate increases. This difference in elevation is called a residual depth anomaly. The thickness of oceanic crust can be calculated from residual depth anomalies and is good to within ±10% of the total crustal thickness [21]. Using these calculations, we can compare the crustal thicknesses of unsubductable oceanic plates with those of plates which are subducting shallowly. This allows us to determine empirically the critical crustal thickness at which a plate becomes unsubductable. We check this empirical result by comparing it to the thickness distribution of unrifted continental crust.

7. Calculating crustal thicknesses in the Ontong Java region

To calculate the crustal thicknesses on the Ontong Java plateau, we must determine the average water depth of the edges of the plateau near the North Solomon trench. We use the methodology of Smith [22,23] to calculate the median water depth within $1/2 \times 1/2°$ squares of the Pacific plate. However, because of the small size and/or topographic variability of areas with thick crust, we explicitly centered the grid elements over them. We then used the median water depth from each square to calculate the residual depth anomaly; that is, the elevation above or below that expected from thermal subsidence and isostasy [24]. Our assumption is that the residual depth anomalies are directly related to the crustal


ONTONG JAVA PLATEAU
NORTH SOLOMON TRENCH
SOUTH SOLOMON TRENCH
WOODLARK BASIN
0
-138
-276
-414
-552

thickness through isostasy [25], according to the formula:

$$t_c = 7.1 + (r_a/218) + 0.000000287 * r_a^2$$

where r_a is the residual depth anomaly in meters, and t_c is crustal thickness in kilometers [26]. This formula is an empirical fit to existing data on crustal thickness, depth versus age, and sediment thickness from the Pacific plate [27].

An assumption implicit in our study is that the crustal thickness of material just entering the subduction zone is the same as the crustal thickness of the oceanic lithosphere that is already in the trench. Because we cannot directly measure the crustal thickness of the subducting slab, we use the residual depth anomaly of material entering the subduction zone to infer the crustal thickness of the oceanic plate within the subduction zone. To infer the crustal thickness of the subducting crust, we selected sites 111–222 km seaward of the trench axis. This distance will sometimes introduce a slight error in crustal thickness (up to 0.5 km), due to the modest height of the flexural bulge, but we feel that this small error is offset by a greater probability of ascertaining the thickness of the oceanic crust within the trench.

The Ontong Java plateau began to subduct southward but was so buoyant that most subduction ceased about 25 Ma ago [19]. Roughly 10 Ma ago a new subduction zone formed to the south, and subduction began with convergence to the north (Fig. 3). The chemical analyses of basalts from the Ontong Java plateau show that they formed within a mantle plume. They were not contaminated by pre-existing continental crust or by oceanic sediment [28]. The entire Ontong Java plateau consists of juvenile oceanic crust.

We infer that the relief and gravity signature of an oceanic trench are directly proportional to the downward force exerted by the negative buoyancy of the subducting slab. On this basis, we believe that only some parts of the Ontong Java plateau are unsubductable. We therefore subdivided the Ontong Java plateau on the basis of trench axis seismicity and gravity anomalies. Regions where there is no trench axis seismicity, we classify as unsubductable (Figs. 3 and 4a). All of the unsubductable Ontong Java lithosphere produces trench axis gravity values of less than −138 mgal (Fig. 4a). We reason that unsubductable oceanic lithosphere is too buoyant to produce a load at the end of the subducting slab. Therefore, the subducting plate will bend only because its lower edge is trapped beneath the former arc lithosphere, and the gravity anomaly in the trench will be less negative. Using these criteria, much of the near-trench portion of the Ontong Java plateau is unsubductable.

Our inferences regarding the buoyant nature of the Ontong Java plateau lithosphere are supported by a comparison of the North Solomon trench with the axial gravity anomaly, plate age, and seismicity patterns of the South Solomon trench (Fig. 4b). The oceanic crust entering the South Solomon trench is much younger than the Cretaceous crust entering the North Solomon trench. All of the lithosphere entering the South Solomon trench has a much higher level of seismic activity than the lithosphere entering the North Solomon trench, consistent with the higher convergence rate along the South Solomon trench. In areas where the subducting crust is older than 5 m.y.,

Fig. 3. Map of gravity and topography in the region of the Ontong Java plateau. Seafloor topography is contoured at 1000 m intervals. Negative gravity anomalies are stippled. The two trenches are delineated by labels. The convergence direction of the North Solomon trench is given to within 15° by the orientation of the compressional axes of 4 shallow earthquakes from the Harvard CMT catalogue (arrows). The North Solomon trench has a small amount of active convergence in areas with shallow (< 50 km) earthquakes (plus signs). Areas of inferred unsubductable crust (squares) have a trench axis with no shallow earthquakes within 50 km and with more positive axial gravity. Areas of subductable crust (hexagons) have an associated trench axis that is seismically active and has more negative axial gravity. The active trench to the south is subducting a spreading ridge in the Woodlark Basin, most of which is younger than 5 m.y. The oceanic plate entering the North Solomon trench is quite old (≥ 120 Ma). Note that both the South Solomon trench near the Woodlark Basin and the North Solomon trench near the Ontong Java plateau have axial gravity anomalies < −138 mgal. Both features involve plates that are very light but the origin of their buoyancy is different. The Woodlark Basin plate is light because it is extremely young, and therefore hot. The Ontong Java plateau plate is light because its mantle has been depleted by melting.

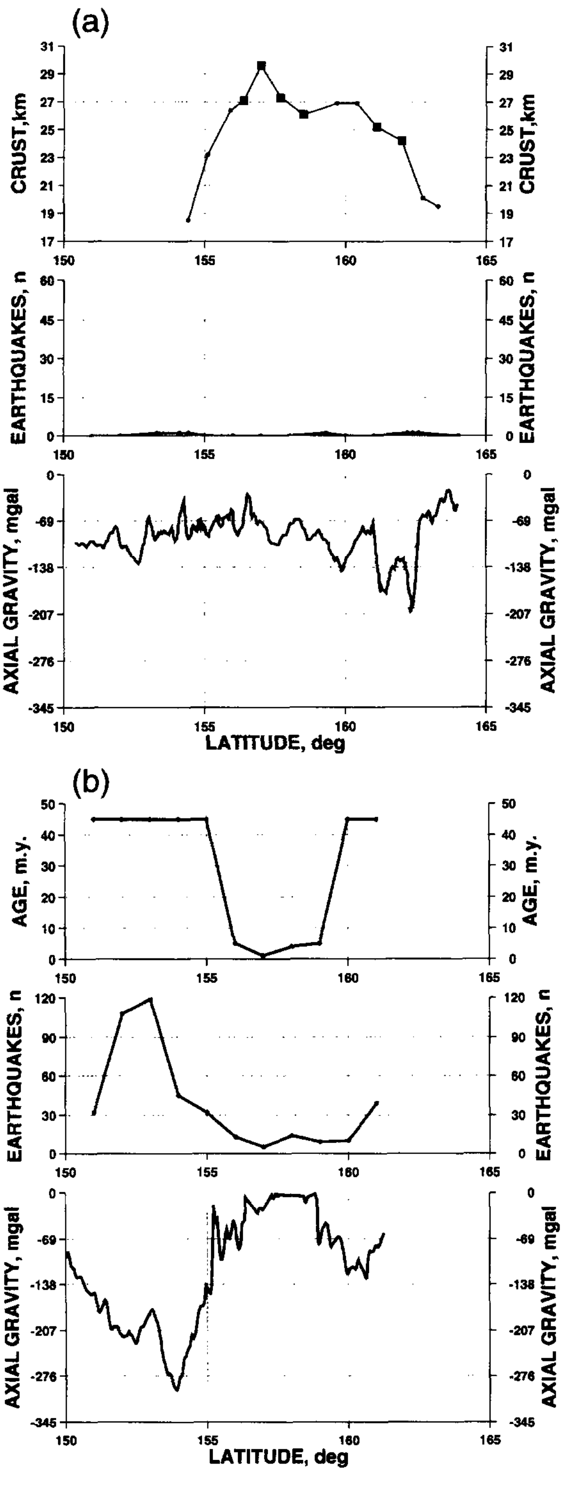


Fig. 4. (a) North Solomon trench and subducting plate: latitudinal variations. Top: Crustal thickness of the subducting Ontong Java plateau. Inferred convergence direction is indicated by arrows in Fig. 3. The convergence direction produces an eastward offset of crustal thickness locations in the top right part of the figure from corresponding locations on the trench axis in the middle and lower parts of the figure. The age (i.e., thermal structure) of the subducting plate does not vary significantly. Squares = thickness of unsubductable crust; dots = thickness of buoyant crust. Note the equivocal interpretation of the three areas where the buoyant crust is over 25 km thick. The two central areas of thick, buoyant crust are defined by one earthquake epicenter. The westernmost area with thick buoyant crust may also be unsubductable, depending upon the exact orientation of convergence. Middle: Shallow earthquakes (less than 50 km focal depth) located within 50 km of the axis of the North Solomon trench. These earthquakes may result from continued convergent activity. Bottom: Minimum gravity value along the axis of the North Solomon trench. (b) South Solomon trench and subducting plate: latitudinal variations. Top: Age (i.e., thermal structure) of the subducting Woodlark Basin plate. The crustal thickness of the subducting plate does not vary significantly. Middle: Number of shallow earthquakes (less than 50 km focal depth) per degree of latitude located within 50 km of the axis of the South Solomon trench. Bottom: Minimum gravity value along the axis of the South Solomon trench. Note that most of the subducting crust over 5 m.y. old produces gravity values along the trench axis that are more negative than − 138 mgal.

the minimum gravity along the axis of the North Solomon trench is much lower than the axial gravity along the South Solomon trench. In the South Solomon trench, the low values of axial gravity north of the Woodlark Basin can be attributed to subduction of oceanic crust formed over the last 5 m.y. Young oceanic crust is quite buoyant, and the low axial gravity in the South Solomon trench north of the Woodlark Basin is consistent with this inference.

8. Results: crustal thickness and subduction style

The results of our compilation of oceanic crustal thicknesses are shown graphically in Fig. 5. The inversion of the residual depth anomalies from the Ontong Java plateau and from areas of shallow subduction shows that oceanic plates with crust from 9 to 27 km thick subduct shallowly and that oceanic plates with crust thicker than 25 km are unsubductable. There is some overlap in Fig. 5. The thickest crust that is actively subducting is 27 km thick, while the thinnest unsubductable crust is 25

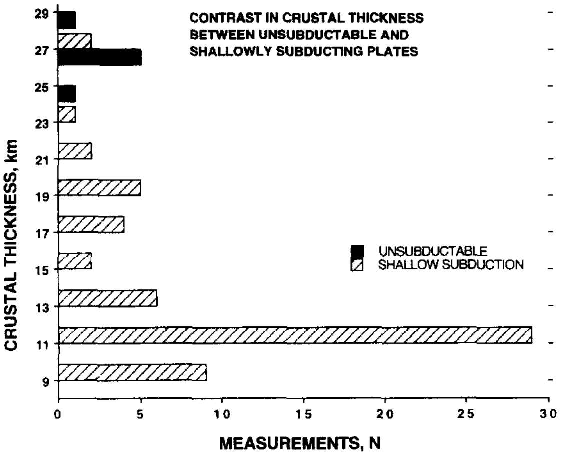


Fig. 5. Histograms of the number of total occurrences versus crustal thickness for unsubductable (black bars) and nearly neutrally buoyant, shallowly subducting (cross-hatched bars) oceanic plates. Although there is some overlap from 25 to 27 km, there is a transition from predominantly shallow subduction to predominantly unsubductable at a crustal thickness of around 25 km. Note that the three sites on the Ontong Java plateau with nearly neutrally buoyant crust and crustal thicknesses of around 27 km could be unsubductable (Figs. 3 and 4a).

km thick. We can therefore conclude that oceanic plates are always unsubductable when their crust is greater than 27 km thick. These apparent differences may be due in part to the 10% calculation error (equivalent to roughly 2 km in crustal thickness) and may be partly due to edge effects.

We can verify if our results are reasonable by comparing the crustal thicknesses we derived for the unsubductable lithosphere in the Ontong Java plateau to the known distribution of continental crustal thicknesses. The rationale is that continental lithosphere is by definition unsubductable. If oceanic lithosphere cannot subduct, it will become a new piece of continent. Thus, the crustal thicknesses of unsubductable oceanic crust should lie within the range of crustal thicknesses of continental crust. In fact, Fig. 6a shows that unsubductable oceanic plates have crustal thickness that overlap the low end of the distribution of continental crustal thicknesses. This suggests that the oceanic crustal layers on accreted plates are thickened after accretion.

We can further verify our estimate of the amount of crustal extraction needed to produce unsubductable lithosphere. We do this by removing all crustal thicknesses in the continental data which are from areas at the ocean–continent boundary or from small ocean basins surrounded by continental crust (e.g., the Black Sea, the Mediterranean Sea). All of these ocean basins are less than 0.2 Ga in age, the typical maximum lifetime of an ocean basin. Thus, it is not certain that these ocean basin plates will remain at the surface of the Earth. We also removed all crust from areas with active rifting (e.g., the Basin and Range province and the Rhine Graben). During continental rifting there is crustal thinning and, therefore, the crustal thickness does not represent its original value.

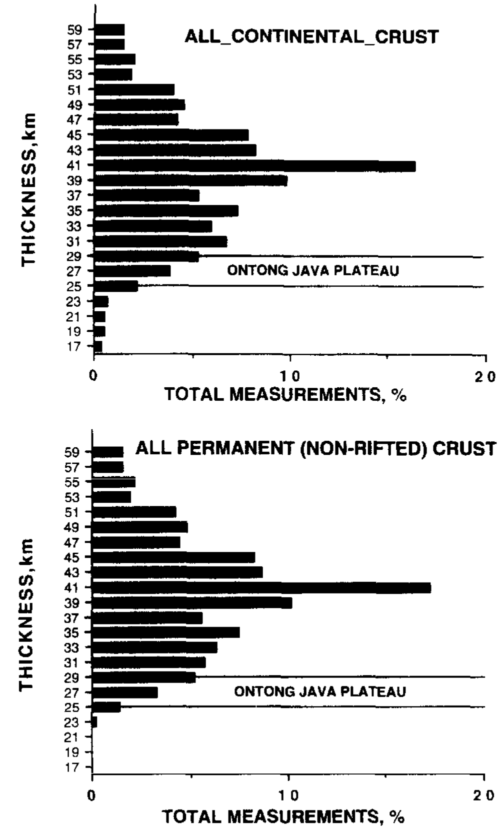


Fig. 6. Above: Comparison of crustal thickness estimates for unsubductable oceanic lithosphere (Ontong Java Plateau) to the crustal thickness distribution of all continental crust. Below: Crustal thickness distribution of all permanent, unrifted continental crust. The line on both figures represents our best estimate from the oceanic data of the crustal thickness dividing unsubductable from shallowly subducting plates.

After we have removed all areas of crust whose thickness is probably not representative of permanent continental crust, we find that the thinnest continental crust is 24 km thick. Twenty four kilometers is very close to the minimum crustal thickness that we obtained from the unsubductable crust of the Ontong Java plateau (25 km) (Figs. 5 and 6b). This is well within the $\pm 10\%$ error of crustal thicknesses derived from seismic refraction (2.4 km). Our analysis supports the results of recent studies which have proposed that some early Precambrian terranes and some Phanerozoic continental terranes originated as oceanic plateaus [29–33].

If oceanic plateaus are accreted to form new continental crust, there must be some recently accreted plateaus whose crustal structure has not yet been altered by collisional or magmatic processes. The Sea of Okhotsk has a thick layer of oceanic sediment, directly underlain by basaltic crust that is 25–30 km thick [34]. The subduction zone in Sakhalin, to the southwest of the Sea of Okhotsk, was active until about 58–60 m.y. ago, based upon K–Ar ages from Sakhalin blueschists [35]. Therefore, the Sea of Okhotsk could represent an accreted oceanic plateau.

The Caribbean plate contains thickened oceanic crust, some of which is 195 Ma in age [36]. The presence of Jurassic age crust in the Caribbean implies that this part of the Caribbean plate formed in the Pacific. Consequently, the eastward dipping subduction zone in Central America must have formed after the Caribbean plate moved east from the proto-Pacific. We suggest that thickened oceanic crust within the western part of the Caribbean Basin produced a polarity reversal of the subduction zone from westward to eastward dipping. Thus, the thickened oceanic crust in the western Caribbean may also be a newly accreted oceanic plateau.

Smaller pieces of recently accreted oceanic crust with mixed alkalic and MORB-like compositions (e.g., hotspot crust) are present in the Marianas arc [37], in western Oregon [38], and in Central America [39]. These smaller pieces of accreted oceanic crust are likely to be quickly overlain by arc volcanics, and will only retain a semi-pristine basaltic composition within the middle to lower crust.

Table 1 contains a list of accreted terranes of Paleozoic and late Precambrian age that include basalts with mixed alkalic and MORB-like compositions (e.g., accreted hotspot crust). Many of these basalts are interbedded with or overlain by basaltic to dacitic island arc volcanics. Table 1 is intended as a supplement to the list of early Precambrian and Mesozoic terranes with hotspot/plateau character-

Table 1
Further examples of accreted terranes with hotspot/plume characteristics

Name	Age	Description	References
Tetagouche Group	Ordovician	MORB-like tholeiitic basalts with interbedded alkali basalts, associated massive sulfides, potassic quartz keratophyre, quartzites and phyllites, felsic metavolcanics	[46] [47]
Gjersvik Terrane	Ordovician	Basalts with MORB and within-plate affinities (e.g., hotspot), associated quartz keratophyres, massive sulfides, graphitic phyllites, ash flow tuffs, arc tholeiitic basalts	[48]
Royvrik Terrane	Ordovician	Tholeiitic and alkaline basalts with a mixed within-plate (hotspot) and mid-ocean ridge affinity, associated massive sulfides, interbedded fine grained marine sediments — unequivocally mid-plate	[48]
Ashland	Vendian to Cambrian	Pillowed low K tholeiitic basalts with alkalic basalts, associated coarse grained gabbro and pyroxenite pods, graphite quartz schist, and garnet mica schist	[49]
Menglian	Carboniferous to Permian	Komatiites with ocean island basalts, radiolarites, phosphorites and carbonates: interpreted as an accreted oceanic plateau or seamount	[50]

istics in [40] and to show that hotspot/plateau crust has accreted throughout Earth history.

9. Discussion

Our results show that a plate does not need a sialic crust to be unsubductable. The variables that determine whether or not a plate will subduct are the degree of partial melting of the source mantle during crustal genesis and the thickness of the layer of depleted lithospheric mantle [14]. Therefore, there are just two requirements for unsubductability: (1) lithospheric mantle depleted by melting which has removed the earlier melting phases and left proportionately more Mg-rich olivine in the residue, and(2) a thick layer of depleted lithospheric mantle. We emphasize that these criteria do not include a sialic crust.

Recent work on the fluid dynamics of melting beneath hotspots has found that, when the plate velocity increases, the degree of partial melting and the thickness of the depleted lithospheric root both increase [41]. Furthermore, the total amount of melting increases with both increasing hotspot temperature and with decreasing thickness (e.g. thermal age) of the overlying plate. That is, the formation of unsubductable material is favored by high velocities of absolute plate motion and by thin pre-existing lithosphere. For example, the basal sequence of the Ontong Java plateau and the accreted seamounts of the Marianas arc formed during the Cretaceous interval of rapid plate motion. These observations support the idea that new continental crust formed from oceanic plateaus and hotspots initially has a basaltic composition.

9.1. *Implications for the origin and evolution of continental crust*

As we discussed earlier in this paper, our model for the incorporation of buoyant oceanic crust into the continental crust does not require that newly formed continental crust have a sialic composition. Rather, we believe that the primordial continental crust more probably had a basaltic composition and became unsubductable because its lithospheric root became buoyant through the depletion of the underlying lithospheric mantle by large degrees of partial melting. This has significant implications for the search for samples of the Earth's most ancient continental crust.

The search for early rocks and minerals has focused mainly on silicic rocks and on zircons, which primarily form within silicic plutons. However, there is no good evidence for an extensive early silicic crust. Instead, the isotopic data appear to require a basaltic crust that remained isolated from the depleted mantle for about 300 Ma [42]. Because the maximum age of subducting Early Archean oceanic plates was < 60 Ma, the average age of subducting oceanic crust was in fact much younger. Therefore, subductable oceanic crust would melt back into the mantle in much less than 60 Ma. The easiest method of isolating a basaltic component from the mantle for ~ 300 Ma is to preserve it as continental crust. That is, the basaltic component which isotopic evidence indicates was first extracted from the mantle and then isolated from the mantle for roughly 300 Ma must have been unsubductable. This implies that the earliest continental crust was most likely basaltic.

Unsubductable oceanic lithosphere within the ocean basins of the Early Archean probably consisted of iron-rich basaltic crust overlying a depleted lithospheric root. It was probably produced through mantle plume activity. Although the oldest continental rocks are most likely basaltic, basic rocks have relatively small proportions of datable zirconium minerals, either baddeleyite (monoclinic ZrO_2), or zircon ($ZrSiO_4$). Consequently, silicic rocks would contribute a disproportionate percentage of the datable zirconium minerals that survived within Early Archean sedimentary sequences. Thus, it is not surprising that the oldest zircons yet found are inferred to have originated within silicic rocks [1].

So far, all of the known examples of the most ancient rocks have been metamorphosed to some degree [2]. Because of their higher melting temperatures, basic rocks are more likely to survive metamorphism than silicic rocks. Many geologists searching for the most ancient rocks have assumed that the basic rocks in a given terrane are either younger or at best contemporaneous with the silicic rocks in the sequence. We suggest that basic rocks from known ancient terranes be routinely dated along with the silicic rocks in order to seek and identify ancient crust.

10. Conclusions

In this paper, we have examined the conditions under which oceanic crust will be unsubductable and be incorporated into a continental land mass. We have shown that the parameters which control lithospheric buoyancy are the degree of depletion in the lithospheric mantle and the thickness of the depleted mantle root. The production of about 25 km of oceanic crust will produce enough depletion in the lithospheric mantle to form an unsubductable root. However, the velocity of the overriding plate, mantle volatile contents, and mantle temperatures are other factors which will influence the relative amount of partial melting and depletion in the lithospheric mantle. For unsubductable plates, such as parts of the Ontong Java plateau (which formed during the Cretaceous episode of rapid plate movement), 25 km of crustal production was sufficient to produce unsubductable oceanic lithosphere. However, during times of slower plate motions, somewhat more crustal production may be necessary for oceanic lithosphere to become unsubductable.

The definition of a continent which we have presented here does not require that it have a silicic crustal composition, but only that it be too buoyant to be subducted back into the mantle. Furthermore, there is no geophysical or geochemical evidence for the existence of an extensive silicic crust during the Early Archean. Consequently, we suggest that Early Archean metabasic rocks (most probably preserved in the middle to lower crust) are the best candidates for the oldest terrestrial rocks.

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