levels in the field-collected colonies were determined using a Carlo Erba NA1500 NCS system. For *Trichodesmium* chlorophyll biomass, the contents of whole 10-l Niskin bottles from stratified depths were gravity filtered onto 5- to 10-μm polycarbonate filters and trichome density determined by direct microscopic enumeration using phycoerythrin epifluorescence. *Trichodesmium* trichome density was converted to chlorophyll terms by a factor derived from direct extraction and determination of chlorophyll per trichome and per colony at each station. *Trichodesmium* biomass was then integrated to the upper 50 m. Standard hydrographic parameters (temperature T, salinity S and density σ_t) were measured by CTD (conductivity–temperature–depth) at each sampling location.

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Preservation of ancient and fertile lithospheric mantle beneath the southwestern United States

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Stable continental regions, free from tectonic activity, are generally found only within ancient cratons—the centres of continents which formed in the Archaean era, 4.0–2.5 Gyr ago. But in the Cordilleran mountain belt of western North America some younger (middle Proterozoic) regions have remained stable[1,2], whereas some older (late Archaean) regions have been tectonically disturbed[1,3], suggesting that age alone does not determine lithospheric strength and crustal stability. Here we report rhenium–osmium isotope and mineral compositions of peridotite xenoliths from two regions of the Cordilleran mountain belt. We found that the younger, undeformed Colorado plateau is underlain by lithospheric mantle that is 'depleted' (deficient in minerals extracted by partial melting of the rock), whereas the older (Archaean), yet deformed, southern Basin and Range province is underlain by 'fertile' lithospheric mantle (not depleted by melt extraction). We suggest that the apparent relationship between composition and lithospheric strength, inferred from different degrees of crustal deformation, occurs because depleted mantle is intrinsically less dense than fertile mantle (due to iron having been lost when melt was extracted from the rock). This allows the depleted mantle to form a thicker thermal boundary layer[4] between the deep convecting mantle and the crust, thus reducing tectonic activity at the surface. The inference that not all Archaean crust developed a strong and thick thermal boundary layer leads to the possibility that such ancient crust may have been overlooked because of its intensive reworking or lost from the geological record owing to preferential recycling.

The North American Cordillera is a broad continental region marked by a long period of tectonic activity, which began in the Palaeozoic era with a series of mountain-forming events and culminated in the Cenozoic era with extension[5]. Deformation appears to be heterogeneously distributed (Fig. 1). The Basin and Range province, which includes much of Nevada and southeastern California, experienced crustal thickening and subsequent large-scale extension (possibly up to 200%)[6]. In contrast, the Colorado plateau, an elevated circular region surrounded on all sides by deformed crust, has remained an island of tectonic quiescence, as evidenced by flat-lying, unfolded and largely unfaulted Palaeozoic sedimentary strata[7].

Given the relative differences in the degree of deformation seen in the overlying crust[5] and the correlation between age and stability observed elsewhere in the continents, the more-tectonized Basin and Range lithosphere might be expected to be younger than that beneath the less-tectonized Colorado plateau. However, Sm–Nd model ages indicate that the crust in the southern Basin and Range (referred to here as Mojavia) is older, formed in Palaeoproterozoic to Archaean times (~2.0–2.6 Gyr ago)[1,2], whereas the Colorado plateau crust formed subsequently in the middle Proterozoic (1.6–2.0 Gyr ago)[1,3]. There are two possible explanations for this unexpected relationship. First, the lithospheric mantle beneath Mojavia may not be as old as the crustal model ages indicate. This might

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occur if the Nd model ages reflect mixing between juvenile middle Proterozoic crust and sedimentary material shed from an adjacent Archaean craton at the time of lithosphere formation[1], or the original ancient lithospheric mantle was removed (as recently suggested for the deep lithosphere of the Sierra Nevada[8]). Second, the lithospheric mantle beneath both regions may be the same age as the overlying crust[9], but the ancient lithospheric mantle beneath Mojavia is inherently weaker than typical cratonic lithospheric mantle owing to a more fertile composition[4,10] and/or the presence of water[11]. We evaluate these two alternatives using Os model ages of peridotite xenoliths recently derived from the lithospheric mantle in both regions.

Assuming that partial melting leads to stabilization of the lithospheric mantle, the Re–Os isotope systematics of peridotite xenoliths (samples of the lithospheric mantle) can be used to date this time of stabilization; this is because partial melting fractionates Re/Os (Re is moderately depleted and Os is sequestered in the residue[12]). As ^{187}Re decays to ^{187}Os (half life of ~42 Gyr), the residue evolves towards unradiogenic ^{187}Os/^{188}Os and diverges from the isotopic trajectory of undepleted convecting mantle[12]. Analogous to Sm–Nd model ages, the time at which melt extraction occurred is determined from the intersection of the two isotopic evolution paths, as constrained by the present-day ^{187}Os/^{188}Os and ^{187}Re/^{188}Os ratios of the sample and the convecting mantle. Because the Os concentration in peridotites is 10–1,000 times greater than that of silicate melts or other metasomatic agents, the Os isotope system in peridotites is considerably more robust to changes imparted by metasomatism and contamination than are isotope systems based on incompatible elements, such as Sm–Nd and Rb–Sr.

The peridotite xenoliths come from the Pliocene Cima volcanic field, located in the Mojavia province, California, and from Miocene-Oligocene minette plugs located in the 'Four corners' region of the Colorado plateau. The Cima peridotites are fresh and devoid of hydrous minerals, while the Colorado plateau samples are partly serpentenized. The Mojavian samples have ^{187}Os/^{188}Os ratios ranging from chondritic to relatively unradiogenic values (see Supplementary Information). The latter reflect long-term isolation in a low Re/Os environment produced by ancient partial melting. The samples show a positive correlation on an isochron plot, with a slope corresponding to an age of 2.5 ± 1.6 Gyr and an initial ^{187}Os/^{188}Os of ~0.112 (Fig. 2a). Using Al_2O_3 as a proxy for the Re/Os ratio[13] results in an initial ^{187}Os/^{188}Os of ~0.113 (Fig. 2b), indicating that depletion in Re and Al probably reflect the same ancient partial melting event. The oldest Re depletion age (assuming Re/Os = 0) for the Mojave samples is 2.4 Gyr (^{187}Os/^{188}Os = 0.1120). Although this represents a minimum age, this sample's ^{187}Os/^{188}Os may closely approximate the 'true' initial ratio because its Al_2O_3 content (0.68 wt%) is below the lower limit (0.7 wt%) at which

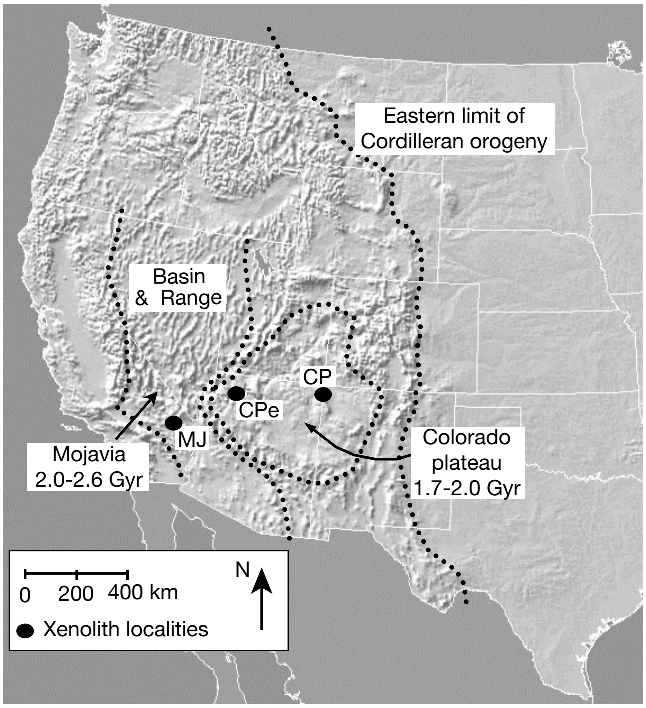

Figure 1 Digital elevation map showing topographic features within the Cordilleran orogenic belt, most of which are associated with late Cenozoic tectonism. The Colorado plateau is an elevated region of mostly undeformed sedimentary strata, surrounded on all sides by regions that have undergone compression and/or extension. Ages represent Sm–Nd crust formation ages[1,2]. Xenolith localities are indicated: CP, central Colorado plateau (The Thumb volcanic dyke); CPe, edge of Colorado plateau (Grand Canyon volcanic field); and MJ, Mojavia (Cima volcanic field).

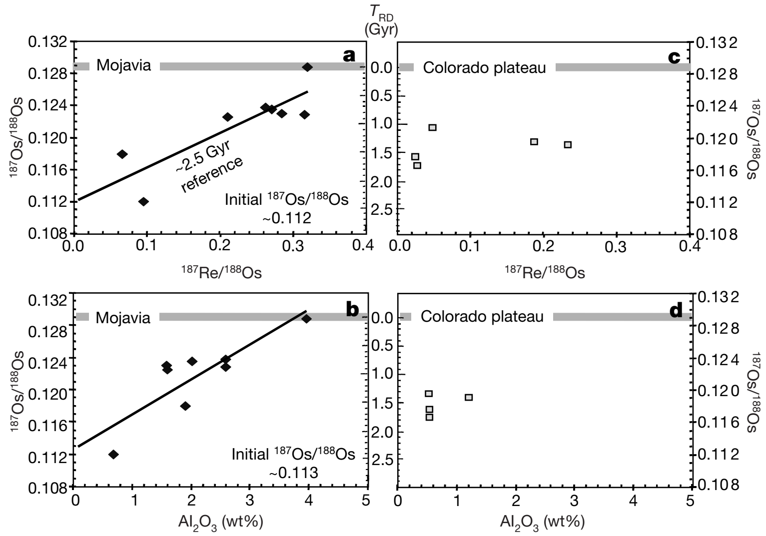

Figure 2 Re–Os isotopic constraints on the age of lithospheric mantle beneath Mojavia and the Colorado plateau. Top panels, Re–Os isotope data (**a**, Mojavia; **c**, Colorado plateau). Bottom panels, Al_2O_3 content (wt%) used as a proxy for Re/Os (**b**, Mojavia; **d**, Colorado plateau). T_{RD} represents Re-depletion model ages, calculated by assuming Re/Os = 0 and extraction from a primitive upper mantle[12] (grey horizontal bar).

Re/Os is suggested to drop to zero during partial melting of the upper mantle[14]. Re–Os model ages on individual samples range between 1.8 and 3.4 Gyr (except for Ki5-139). Collectively, these data point to a late Archaean or earliest Proterozoic age for lithospheric mantle formation beneath Mojavia. This age is consistent with the Sm–Nd model ages of the overlying crust, which thus appear to represent accurately the timing of crust formation[1,2,9].

The Colorado plateau samples (see Supplementary Information) plot along a horizontal array on a Re–Os isochron diagram (Fig. 2c), which does not give any meaningful age significance. Instead, the array is probably due to recent addition or loss of Re (ref. 12). Although the data also do not define a strong correlation between $^{187}Os/^{188}Os$ and Al_2O_3 content (Fig. 2d), the low Al_2O_3 contents and $^{187}Os/^{188}Os$ imply that most of the Re was depleted during a partial melting event, which occurred ~1.6 Gyr ago (Fig. 2c, d), within uncertainty of Sm–Nd model ages of eclogitic xenoliths from the same region[3] and consistent with evidence for old Nd in the lithospheric mantle beneath the Colorado plateau[15,16].

Our data not only confirm that ancient lithospheric mantle persists beneath the southwestern USA despite the protracted Cordilleran orogeny, but also show that the age of the lithospheric mantle beneath each province is indistinguishable from Nd model ages of the overlying crust. Thus continental lithospheric strength is not strictly a function of the formation age of the lithospheric mantle, as might be inferred from the stability of Archaean cratons.

Instead, bulk composition may control the strength of continental lithospheric mantle, by controlling its ultimate thickness. To quantify composition, we use the Mg number (Mg# is the molar Mg/(Mg+Fe) ratio) of a peridotite as a measure of the degree of depletion. This number is an inverse measure of the amount of Fe, and the higher the Mg#, the lower the density. The mode Mg# for the lithospheric mantle of strong Archaean cratons, such as Tanzania[17,18], South Africa[19,20] and Siberia[21], is 0.93, for the weak Palaeoproterozoic to Archaean Mojavia province it is 0.90 (based on compilations in ref. 22 and our own data), and for the strong Proterozoic Colorado plateau it is 0.91 (refs 15, 23) (Fig. 3). Previous studies have shown that most Archaean cratons are underlain by more depleted mantle than post-Archaean regions, leading to an apparent correlation between strength and bulk composition. Our work corroborates these findings by showing that even when age–strength correlations break down, bulk composition still influences lithospheric strength.

Motivated by seismological evidence and the lack of a strong correlation between continents and the long-wavelength geoid, Jordan suggested that continents are underlain by thermal boundary layers, stabilized against convective disruption by compositional buoyancy[4]. This condition, known as the 'isopycnic hypothesis', requires that the negative buoyancy imposed by the colder thermal state of the mantle beneath continents is exactly compensated by a lower intrinsic density of the mantle beneath continents at every level within the viscous thermal boundary layer (but below the elastic mechanical boundary layer). Figure 4 shows isopycnic density curves calculated (see Methods) for various ocean–continent

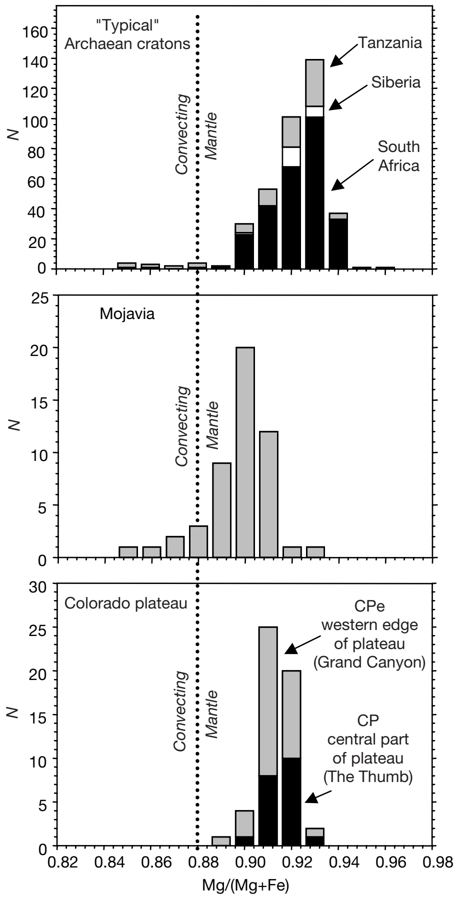

Figure 3 Comparison of the bulk Mg# of lithospheric mantle beneath typical Archaean cratons, the tectonized Mojavia block, and the strong Colorado plateau, *N*, number of samples. Top, a compilation of bulk-rock Mg# for low-temperature peridotites from the Tanzanian[17,18], Siberian[21] and South African cratons[19,20]. High-temperature sheared peridotites were not included in the compilation because these may have been recently refertilized by magmatic processes shortly before eruption. Middle, data from Mojavian xenolith localities based on bulk-rock compilations from ref. 22 and our own data set (olivine compositions). Bottom, data from Colorado plateau samples, which include peridotites from the western edge of the plateau (Grand Canyon[15]) and from central part of the plateau (the 'Four corners' region[10,23]).The estimated Mg# for the convecting upper mantle is taken to be ~0.88 (shown as a vertical dotted line).

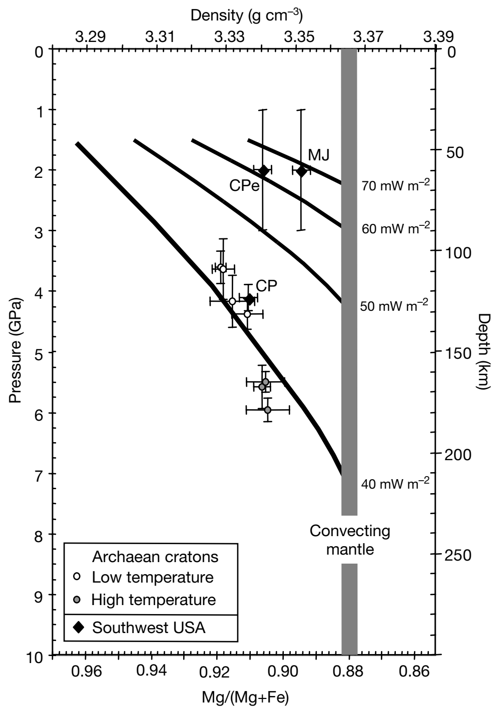

Figure 4 Isopycnic density curves, illustrating the role of bulk composition in determining lithospheric thickness. Curves were calculated for a range of surface heat flows: 70, 60, 50 and 40 mW m^{-2}. MJ, CP and CPe are defined in Fig. 1 legend. Unlabelled points refer to data from different xenolith suites from the Siberian, Tanzanian, and South African cratons (high-temperature peridotites are also shown for reference). Each symbol represents the average density and pressures of all xenoliths within a given xenolith suite. See Methods for details.

temperature differences for a range of surface heat flows and parameters given in the figure legend[24] and in Methods. Although the absolute positions of the isopycnic lines depend strongly on the assumed parameters (namely, crustal heat production and potential temperature of the convecting mantle adiabat), Fig. 4 illustrates that relative differences in bulk composition translate into differences in the thickness of the thermal boundary layer regardless of any uncertainty in the above parameters. The isopycnic hypothesis thus explains the observed correlation between strength and bulk composition because more depleted mantle generates a thicker, colder and hence stronger thermal boundary layer, whereas more-fertile mantle allows for a thinner, hotter and hence weaker thermal boundary layer.

Because the Mojavian peridotites are more fertile than typical cratonic mantle and derive from shallower depths (based on the lack of garnet), they plot on a warmer isopycnic curve ($\sim$70 mW m^{-2}), whereas the xenoliths from the centre of the Colorado plateau (garnet-bearing) plot near the array defined by peridotites from stable Archaean cratons. Ignoring localized regions of high heat flow attributed to recent magmatism, we note that the measured surface heat flow for Mojavia is 60–80 mW m^{-2} and that for the Colorado plateau ranges to lower values (40–80 mW m^{-2})[25]. If all of these regions originally fit the isopycnic condition, as is probably the case, given the lack of large negative geoid or free air gravity anomalies over continents[26], the more-fertile character of the Mojavia mantle requires a thinner thermal boundary layer (50–100 km) than that beneath the more-depleted Colorado plateau and stable Archaean cratons ($\sim$200 km). This is consistent with the lack of garnet in Mojavian xenoliths (implying lithospheric thicknesses less than $\sim$90 km) and the presence of garnet in the Colorado plateau xenoliths. Thermobarometric data on the latter (calculated from the data of ref. 23) show that the Colorado plateau lithosphere extended to at least 120 km depth at the time of xenolith sampling. In this context, we note that spinel peridotite xenoliths from the western edge of the Colorado plateau[10,15] plot along an isopycnic line intermediate between Mojavia and the central Colorado plateau, suggesting that the Colorado plateau lithospheric mantle is thickest at its centre.

The correlation between strength and bulk composition can be explained if the latter dictates the ultimate thickness to which the thermal boundary layer can grow beneath continents, in turn controlling lithospheric strength by defining the thickness of the lithosphere itself. Further thickening of the thermal boundary layer beyond that allowed by the bulk Mg# of the lithospheric mantle would result in this mantle becoming unstable due to its increased negative thermal buoyancy. We suggest that the low viscosity imposed on the residual peridotite by the high temperatures associated with the melting event may have been compensated by a viscosity increase produced by complete dehydration, thus explaining how originally hot, depleted mantle may have eventually stabilized to form thick lithospheric mantle[11,27].

Our findings also have implications for interpreting the ancient rock record. As the Earth may have been hotter and convecting more vigorously in the past, there may have been a secular decrease in the degree of melting, resulting in an apparent correlation between lithospheric mantle composition, and hence lithospheric strength, with age. However, the unusually fertile, ancient lithospheric mantle in the Mojavia province raises the question of how much Archaean lithosphere, characterized by only minor depletion, may have been lost from the geologic record by recycling back into the mantle. If this has occurred, the present distribution of Archaean crust may thus represent a biased sampling of crust underlain by highly depleted mantle (as suggested, for example, in ref. 28). Alternatively, the fact that Mojavian lithospheric mantle still persists despite its fertility may mean that even minor depletion is sufficient to inhibit recycling of continental lithosphere. However, because of its weaker character, such fertile, ancient lithosphere may be more easily overlooked and mistaken for Phanerozoic material on the basis of the degree of deformation or metamorphic age. □

Methods

Calculation of isopycnic density curves

These curves (Fig. 4) were calculated from $\Delta\rho(z) = \rho^0\alpha\Delta T(z)$, where ρ^0 is the normative density (at STP) of the fertile convecting mantle beneath oceans (grey vertical bar), α is the thermal expansion (2.7×10^{-5} per °C), and $\Delta T(z)$ is the continent–ocean temperature difference as a function of depth z (ref. 4). Values of $\Delta T(z)$ were determined by assuming a 1,300 °C potential adiabat for the convecting mantle (adiabatic temperature gradient of 0.5 °C km^{-1}), and continental geotherms with a range of surface heat flows; we assume that heat production of the crust and mantle is 0.5 μW m^{-3} and 0.02 μW m^{-3} (refs 24, 25), respectively. This crustal heat-production value was chosen so that the average pressures and densities of peridotite xenoliths from the stable Archaean Tanzanian, Siberian and South African cratons approximately coincide with an isopycnic curve derived for a surface heat flow of 40 mW m^{-2}, the average measured on stable Archaean cratons[29]. This choice of crustal heat production is within the range allowed by xenolith thermobarometry[24]. Densities were calculated using an empirical correlation between bulk Mg# and density (in g cm^{-3}), $\rho = 4.201 - 0.950\text{Mg\#}$ ($r^2 = 0.74$) from a suite of well-characterized xenoliths from Tanzania[18]. Error bars in Fig. 4 represent the $2\sigma_{mean}$ of the entire population of densities (obtained by propagating the $2\sigma_{mean}$ of Mg#s through the above equation) and pressures for a given xenolith suite (errors in the thermobarometric equation or the equation for calculating density from Mg# are not included). Tanzanian densities were calculated using a linear combination of mineral endmember compositions[30] and their modal abundances (calculated by least-squares regression of whole-rock and mineral compositions). For internal consistency, the density for convecting upper mantle[31] was calculated in the same manner. Pressures for garnet-bearing peridotites (most cratonic peridotites) were calculated using thermobarometers based on the solubility of Al in orthopyroxene coexisting with garnet[32]. For spinel peridotites (for example, MJ and CPe), the pressure was estimated to be between 1 and 3 GPa, based on limits placed on the thickness of the crust and the absence of garnet.

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Supplementary information is available on *Nature*'s World-Wide Web site (http://www.nature.com) or as paper copy from the London editorial office of *Nature*.

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Consistent patterns and the idiosyncratic effects of biodiversity in marine ecosystems

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Revealing the consequences of species extinctions for ecosystem function has been a chief research goal[1–7] and has been accompanied by enthusiastic debate[8–11]. Studies carried out predominantly in terrestrial grassland and soil ecosystems have demonstrated that as the number of species in assembled communities increases, so too do certain ecosystem processes, such as productivity, whereas others such as decomposition can remain unaffected[12]. Diversity can influence aspects of ecosystem function, but questions remain as to how generic the patterns observed are, and whether they are the product of diversity, as such, or of the functional roles and traits that characterize species in ecological systems. Here we demonstrate variable diversity effects for species representative of marine coastal systems at both global and regional scales. We provide evidence for an increase in complementary resource use as diversity increases and show strong evidence for diversity effects in naturally assembled communities at a regional scale. The variability among individual species responses is consistent with a positive but idiosyncratic pattern of ecosystem function with increased diversity.

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We investigated diversity–function relations in mesocosms containing a gradient of species richness, established using intertidal invertebrates. The ecosystem function that we measured was the flux of nutrients (specifically ammonia nitrogen (NH_4-N)) to the overlying water column, a microbial process essential for primary production, which is mediated through the physical working of the sediment by invertebrates[13,14].

Replicate species pools were sampled from locations around the coastlines of northeast Scotland, southwest Sweden and central south Australia. These replicate regional species pools allow us to separate the effects of species richness from any combinatorial effects that might result from species identity. Such between-site comparisons have been made in terrestrial systems[4], and the importance of environment and site effects is well recognized[15–17].

For each site we created a species richness gradient, on the basis of previous knowledge of the local species pools. Species were not selected randomly; rather, they were common species dominating the biomass at each site, and all are known to interact directly with the sediment in which they live. These simple assemblages ranged from one to four species, dependent on locality (see Methods and Supplementary Information). At these low levels of species richness, the effects of diversity are most likely to be manifest. Additionally, species were classified into one of three functional groups so that the effects of functional diversity could also be considered (see Methods). Two treatments were employed: a primary treatment of species

Table 1 Summary of ANCOVA of NH_4-N production across sites and treatments

Source of variation	d.f.	s.s.	m.s.	*F*	*P*
Combined site analysis*					
Site	3	38.108	1.723	158.71	0.001†
Biomass	1	0.937	0.346	31.93	0.001
Site*biomass	3	0.156	0.037	3.44	0.018
Species richness	1	0.058	0.05	4.65	0.032
Error	195	2.12	0.01		
Total	211	43.82			
Individual site analysis‡					
Gullmarsfjord					
Biomass	1	0.234	0.162	50.18	0.001
Species richness	1	0.055	0.017	5.31	0.024
Function richness	1	0.109	0.014	4.49	0.037
Species identity	9	0.206	0.023	7.12	0.001
Error	77	0.248	0.003		
Total	89	0.854			
Ythan a§					
Biomass	1	0.058	0.026	5.18	0.035
Species identity	3	0.067	0.022	4.4	0.016
Error	19	0.096	0.005		
Total	25	0.233			
Ythan b§					
Biomass	1	0.301	0.098	34.09	0.001
Species identity	3	0.438	0.1463	50.92	0.001
Error	40	0.114	0.002		
Total	46	0.917			
Boston bay‖					
Biomass	1	0.828	0.486	25	0.001
Error	39	0.758	0.019		
Total	48	1.732			

d.f., degrees of freedom. s.s., sum of squares. m.s., mean squares.
* Each site was treated as a replicate for the purposes of a large combined analysis. the design treated site as a factor whereas biomass, species richness and functional group richness were treated as covariates. Functional group richness terms were non-significant and are not presented here.
† Significant inter-site differences are responsible for significant species richness terms (two and four species treatments are only present at Gullmarsfjord site, which is significantly different from Ythan and Boston bay sites).
‡ Because of the significant site terms each site was analysed separately. The structure of this analysis regarded species identity (treatment identity) as the main factor, and biomass, species richness and functional group richness as covariates for the purposes of regression analysis.
§ At the Ythan estuary sites, both species richness and functional richness were non-significant (data not shown).
‖ No significant effects were found at Boston bay for species richness, functional richness or species identity.

Supplementary Information

TABLE 1

	rock type†	Al_2O_3	Mg/(Mg+Fe)	Re (ppb)	Os (ppb)	$^{187}Re/^{188}Os$	$^{187}Os/^{188}Os$	T_{MA}‡	T_{RD}‡
Mojavia (Cima)									
Ki5-16	Sp harz	0.68	0.911	0.027	1.35	0.10	0.1120 ± 3	3.13	2.42
Ki5-32	Sp harz	2.60	0.906	0.079	1.42	0.26	0.1237 ± 4	1.97	0.75
Duplicate				0.084	1.66	0.24			
Ki5-110	Sp harz	1.90	0.911	0.021	1.54	0.07	0.1180 ± 2	1.86	1.57
Ki5-139	Pl lherz	3.97	0.891	0.073	1.08	0.32	0.1288 ± 4	0.13	0.03
Duplicate				0.065	1.36	0.23			
Ki5-45	Sp harz	1.59	0.902	0.063	1.04	0.29	0.1230 ± 5	2.63	0.86
CiP98-8	Sp harz	2.03	0.909	0.067	1.17	0.27	0.1235 ± 4	2.16	0.78
Ki5-8	Sp harz	1.70	0.906	0.229					
CiP98-64	Sp harz	1.61	0.908	0.074	1.65	0.21	0.1225 ± 6	1.84	0.92
Ki5-32	Sp harz	2.59	0.903	0.136	2.03	0.32	0.1228 ± 2	3.46	0.87
Colorado Plateau									
Col Plat 120	Gt harz	0.53	0.919	0.025	4.85	0.02	0.1176 ± 2	1.71	1.61
Col Plat 126	Gt harz	0.54	0.915	0.015	2.72	0.03	0.1166 ± 3	1.87	1.75
Col Plat C713	Gt harz			0.023	2.24	0.05	0.1213 ± 3	1.23	1.09
Col Plat 105	Gt harz	1.21	0.921	0.251	5.10	0.23	0.1192 ± 3	3.10	1.40
Col Plat 104‡	Gt harz	0.54	0.918	0.258	6.51	0.19	0.1195 ± 3	2.42	1.35

†sp harz (spinel harz), gt harz (garnet harzburgite), pl lherz (plagioclase lherzolite), cpx (clinopyroxene); Mg#=Mg/(Mg+Fe); ‡ T_{MA} refer to model ages calculated by assuming $^{187}Os/^{188}Os$ of primitive upper mantle is presently 0.129 and $^{187}Re/^{188}Os$ = 0.423 (note that due to problems of late stage Re mobility, T_{MA}'s are likely to represent maximum ages) and T_{RD} refers to Re-depletion model ages calculated assuming $^{187}Re/^{188}Os$ = 0 (therefore they represent minimum model ages); except for Col Plat 104, Re and Os measured on same aliquots; whole-rock powders were made in a metal-free environment using an alumina grinding mill. Samples (1 g aliquots) were simultaneously spiked with ^{190}Os and ^{185}Re tracers; dissolution/spike-sample equilibration were achieved in sealed glass tubes; Os was separated and purified using solvent extraction and microdistillation, and analyzed by negative thermal ionization (Harvard Finnigan Mat 262) using $Ba(OH)_2$ as emission enhancers on Pt filaments (233/236<0.00001); measured ratios were corrected for oxides ($^{17}O/^{16}O$=0.0003708, $^{18}O/^{16}O$=0.002045) and for mass discrimination (exponential law); Re was separated by anion-exchange and analyzed by inductively coupled plasma mass spectrometry; Os and Re process blanks were <3 pg and <6 pg, respectively; external reproducibility of $^{187}Os/^{188}Os$ for 1 ng standard was 3 per mil over course of study.

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Earth science

Hard-cored continents

Andrew A. Nyblade

Each continent contains pockets of ancient crust that appear to have been unaffected by tectonic forces since they formed billions of years ago. Why? There's now a fresh twist on the usual explanation.

The existence of the small, ancient cores of continents, known as cratons, has long been a puzzle. Cratons were created during the Archaean eon, more than 2.5 billion years ago, and form the oldest parts of Earth's tectonic plates. Yet they have somehow remained largely unmodified by tectonic forces. By contrast, younger parts of the continents bear the geological scars \of repeated tectonic buffeting, and appear to be weaker and less stable. So, why are cratons tectonically stable? Lee and co-workers (page 69 of this issue[1]) provide new insight into this question.

Not much is known about the processes that formed cratons in the Archaean. But it has been suspected for some time that their tectonic longevity derives from 'keels' — as on sailing vessels — that extend deep into the Earth (Fig. 1). These keels are made of lithospheric mantle more than 2.5 billion years old and more than 200 km deep. The lithosphere is Earth's outermost rigid layer, and consists of the crust and uppermost mantle. The chemical composition of craton keels is thought to stem from their depletion of the basaltic constituents (Al_2O_3, FeO, CaO) and volatile molecules (H_2O, CO_2) compared with the 'fertile' mantle that is the source of basaltic volcanism along mid-ocean ridges[2,3].

According to theory, a combination of the loss of basalt and volatiles makes the keels strong enough to resist wholesale destruction by tectonic forces. This is because extraction of basaltic constituents during volcanism removes iron from the remaining mantle, making it more buoyant than its surroundings. In addition, removal of volatiles from the mantle during mantle melting increases the melting temperature and stiffness of the remaining material, making it even more resistant to tectonic forces.

Lee *et al.*[1] show that depletion of basaltic constituents does indeed influence the strength of lithospheric mantle, mainly by controlling the thickness to which the keel can grow. But they find that the degree of depletion is not always a function of age. Their evidence is geochemical, and comes from two regions of southwestern United States where small pieces of lithospheric mantle, called xenoliths, have been brought rapidly to the surface by volcanoes. One location is in the southern Basin and Range province, where crust 2.0–2.6 billion years in age is being deformed by tectonic forces. The other is in the Colorado plateau, a tectonically stable region of crust 1.6–2.0 billion years old that borders the Basin and Range province to the east (see map on page 70.)

By measuring the abundance of rhenium and osmium isotopes in the xenoliths, Lee and co-workers show that the lithospheric mantle beneath the sampling localities is similar in age to the overlying crust. The bulk composition of the xenoliths, together with the pressure and temperature conditions under which some of their component minerals formed, also reveal that the lithospheric mantle beneath the Basin and Range province is thinner and less depleted of basaltic constituents — that is, more fertile — than it is beneath the Colorado plateau. So it seems that it is not the age of the lithospheric mantle that correlates with its tectonic stability. Rather, given the greater basaltic depletion and thicker lithospheric mantle found under the younger yet more stable crust of the Colorado plateau, it is depletion and in turn thickness that are the determining factors.

The authors next turn to the question of how this loss of basalt controls the thickness of the lithosphere. Here they draw upon the observation that cratonic keels must in fact be neutrally buoyant, even though they contain less basalt, because they are not associated with significant perturbations in Earth's gravity field. According to the isopycnic (equal-density) condition proposed by Jordan[2], the negative buoyancy resulting

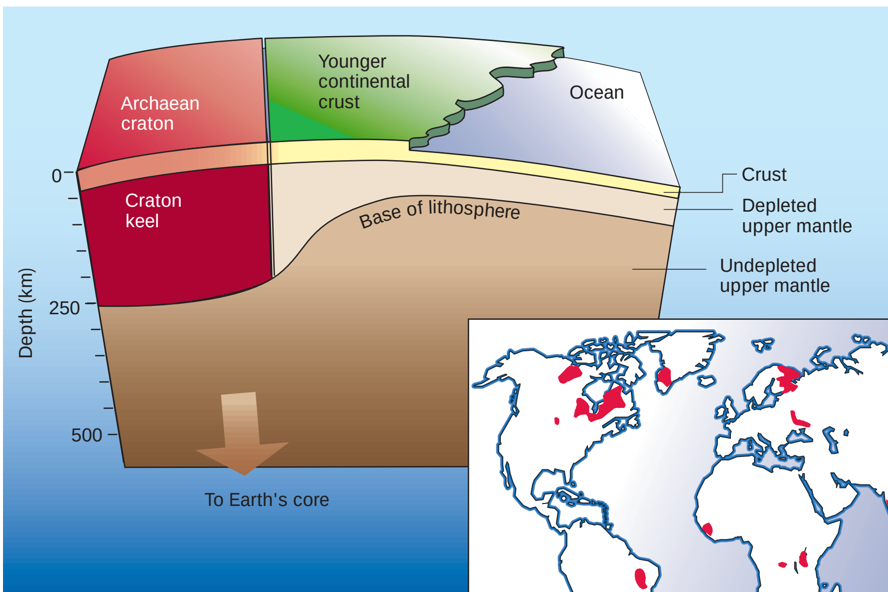

Figure 1 **Archaean cratons and keels. Archaean cratons are tectonically stable continental regions formed before 2.5 billion years ago. Their stability is thought to stem from their underlying keels, which are composed of lithospheric mantle depleted of basaltic components and are at least twice as thick as the lithospheric mantle beneath younger parts of the continents and oceans. As reported by Lee *et al.*[1], the thickness of the keel is controlled by the degree of basalt depletion in the lithospheric mantle. The inset map shows the global distribution of Archaean cratons. (Main graphic modified from ref. 2.)**

from basalt depletion is offset by the positive buoyancy that results from the lower temperatures in the keel relative to those in the surrounding, convecting mantle.

Lee *et al.*[1] compare estimated variations in the thickness, temperature and density of the lithospheric mantle with values calculated by assuming the isopycnic condition, and find that the condition holds for the southwestern United States. So, if the lithospheric mantle were to thicken further by conductive cooling in their study area, the deepest parts of the mantle lithosphere would become less buoyant and sink into the convecting mantle — as long as chemical depletion did not offset the increase in negative thermal buoyancy. In this way, the degree of basalt depletion modulates the thickness of the lithospheric mantle.

A further implication of the results[1] is that more continental crust may have formed before 2.5 billion years ago than is indicated by the present distribution of cratons. The presence of thin Archaean lithospheric mantle beneath the Basin and Range province raises the possibility that there could be similar lithospheric mantle under other regions of the continents. We simply may not have recognized it because Archaean lithospheric mantle is commonly, but mistakenly, assumed to be thick and strong. Alternatively, the weakness of this thin lithosphere may have led to its preferential destruction through tectonic recycling. In either case, much more continental crust could have formed in the Archaean than today's distribution of cratons would indicate. ■

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Fundamental physics

Resistance of a perfect wire

Albert M. Chang

Intuition tells us that a wire without defects should have zero resistance. But in the real world all conductors, however perfect, have some resistance. A new study confirms that electrical contacts are the problem.

It is the dream of physicists and electrical engineers alike to build electronic devices that can conduct electrical currents with the minimal amount of resistance. The miniaturization of electronics will soon lead to devices of the smallest possible physical dimensions, in which quantum effects become important. In this context, the ultimate conductor would be a very thin one-dimensional wire that has no defects to inhibit resistance-free currents. Electrons in such a wire are ballistic — that is, the wire is so clean that the distance travelled by the electrons between collisions is longer than the wire itself. On page 51 of this issue de Picciotto *et al.*[1] describe electrical conduction in a nearly perfect, ballistic one-dimensional wire. This groundbreaking work helps to establish fundamental limits on the current-carrying capacity of ideal wires and their connections.

Imagine a perfect, extremely thin, straight wire in which electrons are allowed to move only along the wire. A perfect wire has no defects, kinks or obstacles other than a connection at each end to allow current to pass through an external circuit, and perhaps two probes along the wire to measure the voltage (Fig. 1a, overleaf). Will the motion of the electrons and hence conduction of electricity in this wire proceed without resistance? De Picciotto *et al.*[1] have created this imaginary wire in the laboratory. They find that the resistance of a wire can be separated into two parts: an 'intrinsic resistance' due to the scattering of electrons by imperfections in the wire, and a 'contact resistance' associated with the connections to the external circuit. The intrinsic resistance is measured by the two voltage probes, which draw negligible current. The authors find that in their defect-free wire the intrinsic resistance does actually reach zero, although there is a finite contact resistance of around 13 kΩ.

The vanishing of the intrinsic resistance agrees with the simple notion that the current-carrying electrons should move freely if there are no obstacles. Our picture of electrical resistance as the result of momentum-changing deflections on charge-carriers dates back to the work of Drude around 1900. The most effective deflections are those that scatter charge-carriers in the opposite direction to that of their flow. In the late 1950s, Landauer[2] became fascinated with the idea of electronic miniaturization and proposed a conceptual framework for understanding electrical conduction in one-dimensional wires. Landauer realized that the wire can be thought of as being connected to electrochemical potential reservoirs, in which many electrons of different energies are available for conduction (Fig. 1b). Once an electron enters the wire it cannot change energy, and only momentum changes can affect the electrical current. So where the wire is connected to the reservoirs there is a mismatch of energies and a contact resistance develops. Conversely, in a regular three-dimensional wire, the contact resistance is very small and so tends to go unnoticed. But in a one-dimensional conductor, with no impurities, the intrinsic resistance must vanish although the contract resistance is high.

Until the 1980s most current measurements on mesoscopic devices (typically a micrometre or less in size) used a two-terminal geometry, in which voltage difference is measured solely between the current source and drain. But after Webb *et al.*[3] developed the technology to place multiple terminals on small metallic wires and rings, it was discovered that the four-terminal — two voltage probes in addition to two current leads — resistance of a non-ballistic wire changes if the direction of an external magnetic field is reversed. This unexpected result was explained by Buttiker[4], who generalized Landauer's ideas to devices with multiple terminals. He pointed out that, in the absence of magnetic impurities, a four-terminal resistance should be unchanged by swapping over the current and voltage probes and reversing the magnetic field.

After considering the multiple-terminal case, Buttiker proposed that a two-terminal resistance must always contain a contact resistance, even if the wire is ballistic. This contact resistance can be measured directly between point contacts that are shorter than the mean free path of electrons in a non-ballistic conductor[5,6]. Two-terminal measurements of ballistic wires also found their resistance to be quite large, around 13 kΩ. Is this resistance just the sum of the contact resistances?

De Picciotto *et al.*[1] have built a multi-terminal one-dimensional conductor to address this issue. Classically a wire is considered to be one-dimensional if its width exactly accommodates the size of the charge carriers with no room for wiggling. Electrons are essentially point objects, so this is technically unfeasible. But this is where quantum mechanics comes to the rescue. If an electron is made to occupy the lowest quantum-mechanical energy state in the lateral directions, without access to higher excited states, it would be free to move only in one dimension. The key is to confine electrons so tightly that the energy levels of the excited states are too high for them to reach. One way of doing this is to cool the wire to sufficiently low temperatures — but not superconducting temperatures — so that the excited states are thermally inaccessible and conduction is strictly one dimensional.

To create the perfect one-dimensional