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Making and altering the crust: A global perspective on crustal structure and evolution

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ABSTRACT

Crustal structure preserves a unique record of physical and chemical conditions of its formation and later modification by geodynamic processes. The existence of broad global correlations between crustal structure and tectonic settings led to models of crustal typization by 1D crustal columns based on absolute thicknesses of crustal layers and the Moho depth.

Here we propose a fundamentally different approach to typify the crust and geodynamic models of crustal evolution. We demonstrate that the relative ratio of the thicknesses of three principal crustal layers (sedimentary/felsic-intermediate/mafic in continents and Layer1/Layer2/Layer3 in oceans) is a fundamental characteristic of the crust. The relative ratio uniquely specifies the crustal structure in different tectonic settings and is independent of the absolute values of thickness of the crustal layers and the Moho depth. We analyze this new fundamental characteristic of the crust by ternary diagrams based on seismic models for continental and oceanic crustal structure in the northern Eurasia – northern Atlantics region and for selected oceanic provinces of different geodynamic origin, where seismic models for the crust are available. We present global and regional trends of crustal evolution and, as a practical application of the new approach, calculate average crustal densities in different continental and oceanic tectonic settings. These values range from ca. 2700 kg/m^3 in deep basins, to 2775 kg/m^3 in orogens and shelves, 2800 kg/m^3 in rifts and some ocean hotspots, 2800–2850 kg/m^3 in shields and platforms, 2900 kg/m^3 in back-arc basins and aseismic ridges, and may reach 2950 kg/m^3 in the Pacific hotspots.

1. Introduction

The existence of two fundamental crustal types, oceanic and continental, was recognized from earthquake studies by Gutenberg in 1924. The principal differences are in composition and thickness: oceanic crust lacks the felsic-intermediate layer and is significantly thinner than continental crust. Once the crust is formed, it undergoes a series of mechanical and chemical modifications which change its total thickness, composition, and thickness of individual crustal layers. The extent and type of these modifications depend on tectonic and geodynamic settings.

Oceanic crust is formed under similar conditions at mid-ocean ridges and has a globally uniform thickness of ca. 7–8 km (White et al., 1992), with a slightly thinner crust formed at ultra-slow spreading ridges (Small, 1998). Further transformations of oceanic crust are limited chiefly to magmatic modification, such as at hot spots, and to sediment deposition, particularly in old oceans and along continental margins, in both cases leading to thickening of oceanic crust.

The formation of continental crust remains enigmatic. Generation of the granitic upper layer requires water (Wedepohl, 1995), and it is commonly accepted that it is formed primarily at volcanic island arcs (Foley et al., 2002). However, the details of the process (Wedepohl, 1995; O'Neil and Carlson, 2017) as well as transformation of arc crust into continental crust (Kelemen and Behn, 2016) are still debated, and composition of the lower crust remains controversial (Rudnick and Fountain, 1995). Large variability of thermo-chemical conditions at mantle wedges leads to a significant diversity in the structure of continental crust when it is formed, as observed in modern island arcs (e.g. Iwasaki et al., 2013).

A long tectonic life of continental crust leads to its significant reworking by plate tectonics processes and crust-mantle interaction (Brown and Rushmer, 2006; Hawkesworth and Kemp, 2006; Artemieva and Meissner, 2012). This includes, among other processes, mechanical extension (Ruppel, 1995), delamination (Menard and Molnar, 1988; Kay and Kay, 1993), relamination (Kelemen and Behn, 2016), magmatic intra- and underplating

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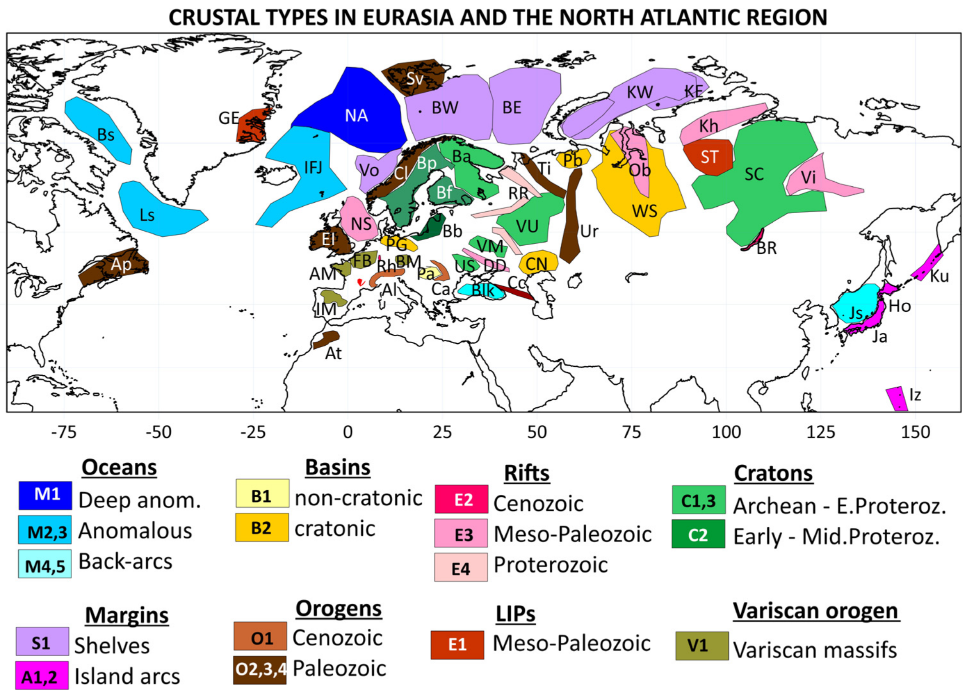

Fig. 1. Crustal types in the northern Eurasia and the North Atlantic region. Oceanic provinces of different tectonic origin (not shown on the map) are also included into the analysis. The two-letter name abbreviation for different tectonic provinces is explained in Table S1. The same color code as in the map is used in Figs. 2, 4 and 7. (For interpretation of the colors in the figure(s), the reader is referred to the web version of this article.)

(Voshage et al., 1990; Thybo and Artemieva, 2013), metamorphic reactions (Rudnick and Fountain, 1995; Austerheim et al., 1997), sedimentation and erosion (Ziegler and Cloetingh, 2004). As result, thickness of the entire crust and thicknesses of its internal layers may change significantly. In extreme cases, some crustal layers can be entirely missing, as for example in Variscan Western Europe the lower crust is nearly absent (Aichroth et al., 1992).

Systematic studies of the crust based on controlled-source seismology led to recognition of broad correlations between crustal structure, crustal age and tectonic settings (Meissner, 1986; Beloussov et al., 1992). These studies focused on the absolute thicknesses of major crustal layers, as recognized by seismic velocities, and the total crustal thickness, and has led to classification of crustal models, which has formed basis for global crustal models starting with CRUST5.1 (Mooney et al., 1998).

Importantly, the role of relative contributions of the major crustal layers has often been overlooked, and until now, the total crustal thickness is considered as an important parameter in crustal classification. Yet a regional analysis of the relative thicknesses of just two crustal layers demonstrated that they provide an efficient indicator of crustal tectonic origin and an extent of crustal reworking (Artemieva and Thybo, 2013): in Europe, the thickness of the granitic-intermediate layer normalized by the thickness of the crystalline basement is <0.3 for oceanic crust, 0.3 to 0.5 for transitional crust, 0.5 to 0.7–0.8 for crust of stable platforms, and >0.8 for extended continental crust.

Here we analyze the role of three major crustal layers, including sediments, in defining uniquely the crustal type and discuss major patterns in the crustal structure in different tectonic settings in the northern Eurasia – North Atlantic region. We take advantage of two recent crustal models for Eurasia, EUNAseis (Artemieva and Thybo, 2013) and SibCrust (Cherepanova et al., 2013), which are constrained by all available seismic data on the crustal structure for the region which extends from the Eastern Canada (80°W) to Siberia (140°E) (Fig. 1).

To extend the list of different tectonic settings covered by the analysis, we also include several oceanic regions in the Pacific and Indian oceans where seismic data on the crustal structure is available: the Japan–Kurils and the Bonin arcs (Iwasaki et al., 2013), the Lau Basin (Crawford et al., 2003), the Cocos Ridge (Walther, 2003), the Hawaii hotspot (Hill, 1969), and the Louisville Seamount (Contreras-Reyes et al., 2010) in the Pacific ocean; the Reunion (Gallart et al., 1999) and the Laccadive (Gupta et al., 2010) volcanic islands, the Laxmi (Naini and Talwani, 1982) and the Ninety-East (Grevemeyer et al., 2001) ridges, all in the Indian Ocean, and the Walvis ridge (Fromm et al., 2015) in the South Atlantic ocean.

We demonstrate that for each tectonic setting the internal crustal structure is fully specified by the relative thicknesses of three major crustal layers, which therefore are the major indicators of the crustal geodynamic type. The new approach is completely independent of the crustal thickness.

2. Tectonic provinces

Tectonic provinces (ca. 70 in total, Table S1) are selected to represent as many different tectonic settings as possible in continental and oceanic domains of northern Eurasia and in oceans globally. Given the uneven data coverage for the seismic structure of the crust in Eurasia (see Artemieva and Thybo, 2013; Cherepanova et al., 2013), only regions with a high density of seismic profiles, and therefore with a well-known crustal structure are included into the analysis (Fig. 1). Major tectonic types are further subdivided into a number of sub-types (24 in total, Table 1), depending on geodynamic settings. For some structures (e.g. the Black Sea) the tectonic classification is non-unique. We note that our results and conclusions are independent of the choice of the tectonic type, and instead, they provide a new basis for tectonic regionalization based on the fundamental differences in the internal crustal structure in different geodynamic settings.

The following tectonic provinces were selected for the crustal structure analysis (Table S1, Fig. 1):

Table 1
Normalized thickness of three major crustal layers in different crustal types.

Tectonic province	Type	Normalized by total crustal thickness			Crustal thickness (km)
		Sediments/Layer 1	Felsic-interm. layer/Layer 2	Mafic layer/Layer 3	
Cratons	C1–C3	0.05 ± 0.03	0.60 ± 0.03	0.35 ± 0.05	44.9 ± 3.5
Basins Cz	B1	0.24 ± 0.07	0.61 ± 0.03	0.15 ± 0.03	30.5 ± 2.1
Basins Mz–Pz	B2	0.24 ± 0.10	0.47 ± 0.13	0.29 ± 0.11	40.3 ± 1.7
Orogens Cz	O1	0.11 ± 0.03	0.73 ± 0.09	0.16 ± 0.09	41.7 ± 4.2
Orogens Pz	O2	0.07 ± 0.02	0.67 ± 0.05	0.26 ± 0.07	35.4 ± 3.2
Variscan massifs	V1	0.07 ± 0.04	0.86 ± 0.07	0.07 ± 0.04	33.5 ± 1.3
Rifts Cz and Pz	E2–E3	0.17 ± 0.04	0.48 ± 0.05	0.35 ± 0.04	38.2 ± 5.3
Dnieper–Donets rift (axial part)[a]	E2	0.52	0.15	0.33	39
Arctic shelves	S1–S2	0.22 ± 0.04	0.48 ± 0.08	0.30 ± 0.07	33.0 ± 2.8
Volcanic arc (Japan–Kurils)[b]	A1	0.04 ± 0.02	0.46 ± 0.02	0.50 ± 0.01	33.8 ± 0.4
Volcanic arc (Honsu)[a]	A1[a]	0.03	0.74	0.23	35
Normal ocean, no sediments[a]	M1o	0	0.38	0.62	8
Normal ocean, 3 km sediments[a]	M1	0.27	0.28	0.45	11
Labrador Sea/Baffin Bay	M2	0.27 ± 0.03	0.40 ± 0.04	0.33 ± 0.06	15.0 ± 1.4
Anomalous North Atlantics	M3	0.15 ± 0.02	0.40 ± 0.01	0.45 ± 0.01	20.0 ± 2.8
Black Sea[a]	M4	0.36	0.43	0.21	28
Back arc basins (center)	M5	0.11 ± 0.04	0.24 ± 0.10	0.65 ± 0.10	10.9 ± 5.1
Hotspots (Indian Ocean)	H1	0.36 ± 0.06	0.19 ± 0.04	0.46 ± 0.04	19.2 ± 4.7
Hotspots (Pacific Ocean)	H2	0.12 ± 0.06	0.05 ± 0.01	0.83 ± 0.06	15.8 ± 4.6
Aseismic ocean ridges	H3	0.11 ± 0.03	0.17 ± 0.07	0.73 ± 0.08	23.6 ± 8.0

[a] Local data.
[b] Honshu excluded.

- Precambrian cratons (types C1–C3), including shields and platforms (Baltic Shield, Russian platform, Ukrainian Shield, Voronezh Massif, Siberian craton);
- sedimentary basins (types B1–B2), including Cenozoic (Pannonian) and Meso-Paleozoic basins (Polish–German, North Caspian, Pechora, West Siberia);
- orogens (types O1–O4), including Cenozoic (Alps, Caucasus, Carpathians) and Paleozoic (North Appalachians, Norwegian Caledonides, Caledonides of UK and Ireland, Timan ridge, Urals, Anti-Atlas/Atlas mountains, Svalbard) orogens;
- Variscan orogen (type V1), including the Gondwana massifs (Iberian, Bohemian, Armorican, Brabant);
- large igneous provinces (LIPs) (type E1), including Paleozoic (the Siberian LIP) and Mesozoic (the North Atlantic Igneous Province in Eastern Greenland);
- extended continental crust (types E2–E4), including active rifts (Rhine Graben, Baikal), Meso-Paleozoic paleorifts (the Central Graben of the North Sea, Oslo and Dnieper–Donets rifts in Europe, and Ob, Khatanga and Viluy rifts in Siberia), and Proterozoic rifts (aulacogens) of the East European Craton;
- continental shelves and margins (type S1), including shelves of the Arctic Ocean (Barents and Kara) and the Voring margin of the North Atlantic Ocean (off-shore Norway);
- oceans (types M1–M3), including "normal" oceanic crust (with differing thickness of sediments) and anomalous oceanic crust (ocean plateau) that does not fit the age-bathymetry predictions (the Labrador and Baffin seas, the North Atlantic Ocean around Iceland, the Jan Mayen block, and the Iceland–Faroe region);
- off-shore back-arc basins (types M4–M5), including Western Pacific (the Japan Sea and Lau Basin) and the Black Sea; the latter may have been formed as a Cretaceous back-arc basin (e.g. Zonenshain and Pichon, 1986) and its crustal structure is not well constrained by seismic studies; the back-arcs of the Mediterranean are excluded because of their small size and the lack of seismic data on the inner structure of their crust;
- volcanic island arcs (type A1), including the Kurils, Japan, and the Izu–Bonin arcs;
- ocean hotspots and volcanic provinces (types H1–H2), including the Hawaii and Louisville hotspots in the Pacific Ocean and the Reunion, Laccadive and Laxmi volcanic provinces in the Indian Ocean;
- aseismic ocean ridges (type H3), including the Cocos, Walwis, Bonin, and the Ninety-East ridges.

3. Analysis of crustal structure

Our analysis is based on published seismic data on the inner structure of the crust for different tectonic provinces of the northern Eurasia. The crustal models EUNAseis and SibCrust are constrained at a lateral spacing of at least 50 km and in more detail where the crustal structure changes sharply (for details see Artemieva and Thybo, 2013; Cherepanova et al., 2013). Following a conventional approach (e.g. Christensen and Mooney, 1995), in general the layers of the continental crust (sediments, upper crust, middle crust, lower crust and high-Vp lowermost crustal layer) are defined by Vp seismic velocities of <5.8 km/s, 5.8–6.4 km/s, 6.4–6.8 km/s, 6.8–7.2 km/s, and >7.2 km/s, respectively. However, when seismic surveys included reflection data and other geophysical information, the boundaries between the crustal layers were accepted as interpreted in the original publications.

We sampled the gridded merged crustal database on a regular 0.5 deg × 0.5 deg grid. The parameters characterizing the crustal structure (the thickness of different crustal layers and the total crustal thickness) are summarized in Table 1 for different tectonic provinces (Table S1 provides details for the individual crustal provinces). In regions where the crustal structure is strongly heterogeneous (e.g. rifts and back-arc basins), local crustal data along seismic profiles are also included into the analysis (cf. Artemieva and Thybo, 2013; Cherepanova et al., 2013) (location names marked by * in Fig. 2).

For oceanic provinces, not covered by the EUNAseis and SibCrust seismic crustal models, we used local seismic profiles, where the oceanic layers are recognized by seismic Vp velocities, similar to the continental crust. Note that our data are dominated by continental crust, and hereafter we use the continental terminology for basement layers. For oceanic crust, the three layers that we discuss correspond to layers L1 (sediments), L2 (mostly basalts), and L3 (gabbro).

4. Results: global crustal structure styles

We present the results as ternary diagrams, where the total crustal thickness (100%) sums up from thicknesses of 3 crustal lay-

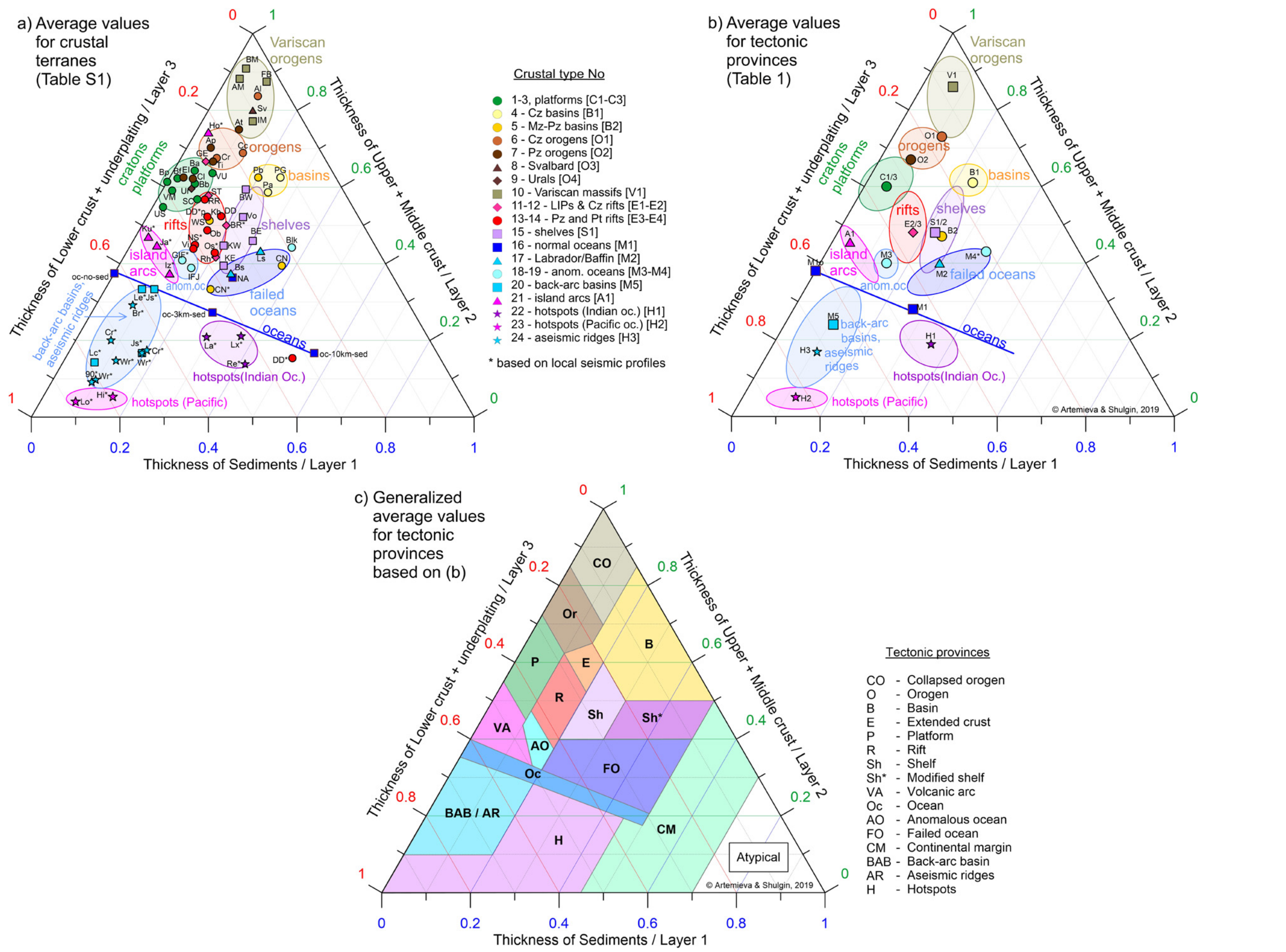

Fig. 2. Ternary diagram for relative thicknesses of three major crustal layers in continental and oceanic crustal provinces: (1) sedimentary cover (or Layer 1), (2) upper-middle crust of felsic-intermediate composition (or Layer 2 for oceanic crust), (3) lower and high-Vp lowermost crust of mafic composition (Layer 3 for oceanic crust). (a) Average values for crustal terranes. Data are based on a $30' \times 30'$ lateral sampling, except for locations (marked by *) with strong crustal heterogeneity where local seismic data from the central part of a structure are used. Color code for crustal provinces is the same as in Fig. 1. Colored ovals show typical ranges of the values including the uncertainties (Table 1). (b) Relative thicknesses of three major crustal layers averaged for different tectonic types. (c) Generalized values for tectonic provinces.

ers (Fig. 2a): sediments (or L1 for oceans); the felsic-intermediate upper and middle crust (or L2 for oceans); the mafic lower and lowermost crust (or L3 for oceans). The choice of these three crustal layers reflects major processes of crustal formation and modification: (i) thickness of sedimentary cover reflects the subsidence/erosion history; (ii) the presence of the felsic-intermediate crust is a characteristic feature of continental crust; (iii) crustal tectono-magmatic modification often leads to a significant alteration of the lower crustal layer, either by its delamination as in most of the Western Europe (Menard and Molnar, 1988), or by magmatic underplating and intrusions as observed globally in various continental and oceanic settings (Thybo and Artemieva, 2013).

The major result of our analysis is that all known crustal types both in oceanic and continental settings are fully – and uniquely – described solely by the ratios of the thicknesses of the three principal crustal layers (Fig. 2). Ternary diagrams with other combinations of the crustal layers do not show any systematic patterns. Importantly and surprisingly, the total crustal thickness is of no significance: these are only the relative thicknesses of the three major crustal layers that are important. When typical crustal columns are plotted conventionally (Fig. 3), no systematic patterns nor trends can be recognized easily (Fig. 3b), even when the values are normalized by the total crustal thickness (Fig. 3a).

Relative thicknesses of the felsic-intermediate crustal layer and the mafic layer are well correlated (Fig. 4c). This suggests that processes responsible for crustal formation and modification affect both of the crystalline layers. Our results extend conclusions of the previous study, where the importance of the relative ratio of the felsic-intermediate and mafic layers in determining crustal tectonic and geodynamic origin has been recognized for the European region (Artemieva and Thybo, 2013).

The new approach to crustal classification provides an understanding of principal trends of crustal evolution both in continents and oceans (Fig. 5). This includes the fundamental trends of granitization and formation of continental crust, orogenic mountain-building and orogenic collapse, including lower crustal delamination, basin formation by crustal extension and thermal subsidence, oceanization of continental crust, and magmatic intrusions that may convert basins to shelves and may create ocean hotspots and aseismic ridges. All of these crustal modification trends cause a significant redistribution of crustal material between the three major crustal layers, while the corresponding changes in the crustal thickness are secondary and do not describe in full processes of crustal evolution: clearly, one parameter (crustal thickness) cannot specify processes which are specified by three parameters (Fig. 5).

Practical applications of our findings include several aspects.

1) The unique relative fractions of the three major crustal layers allow for identification of possible mechanisms of transformation of one crustal type to another one. Importantly, we analyze changes in relative fractions of different crustal layers, not in the absolute values. For example, a relative increase in thickness of the granitic layer implies either its absolute growth, e.g. by production

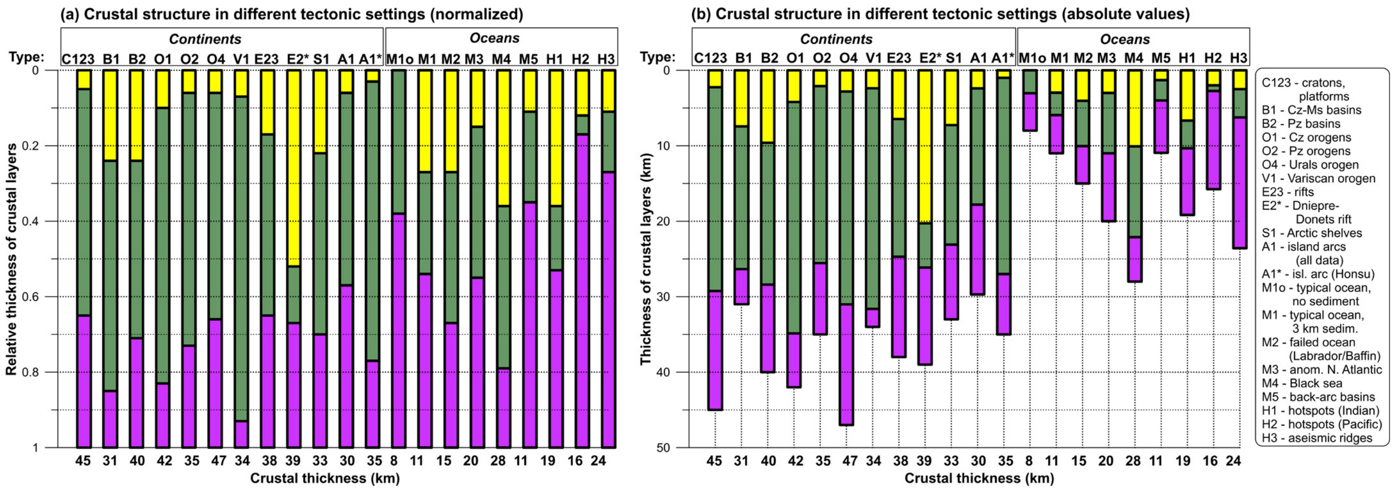

Fig. 3. Relative (a) and absolute (b) thickness of three major crustal layers (sediments or Layer 1; upper and middle crust or Layer 2; lower and lowermost crust or Layer 3) in different tectonic provinces worldwide (crustal type is marked above each section) (see Table 1 for details). In (a), thickness of the crustal layers is normalized by total crustal thickness (numbers below the sections). Crustal data are based on a 30′ × 30′ lateral sampling, except for locations (marked by *) with strong crustal heterogeneity where local seismic data from the central part of a structure are used.

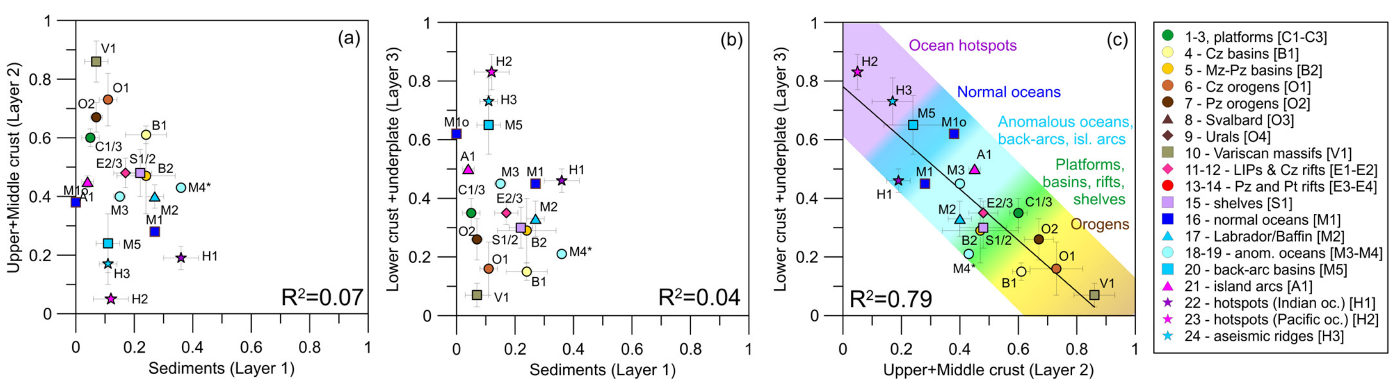

Fig. 4. Cross-plots for relative thickness of three major crustal layers in different tectonic provinces worldwide. The two-letter name abbreviation for tectonic provinces is explained in Table S1. Color code is the same as in Figs. 1, 2. A strong correlation between relative thicknesses of the felsic-intermediate and mafic layers indicates that both layers are affected by major processes of crustal formation and modification.

of new continental crust in volcanic arcs, or a significant thinning of the lower crustal layer, e.g. through delamination. Likewise, a relative increase in thickness of the lower crust can take place through basaltic additions during magmatic events, or by thinning of the upper crust e.g. by extension, with both processes leading to a decrease in the relative ratio of the upper crust.

2) Relative fractions of three major crustal layers uniquely specify crustal structure at different geodynamic settings (Fig. 2b). It means, for example, that the tectonic origin of the crust from a region with the unknown geology (e.g. ice-covered Greenland and Antarctica) can be identified from thicknesses of three major crustal layers when a detailed seismic image of the crust is available.

3) Each crustal layer has a characteristic range of densities (and other physical properties). The unique relative fractions of three major crustal layers in different tectonic settings provide basis for calculation of the typical average density of the crust formed by different tectonic regimes.

5. Discussion

5.1. Regional trends

We next discuss major trends in crustal evolution for crustal provinces of northern Eurasia and northern Atlantics, and start from volcanic arcs, the tectonic environment where the continental crust is primarily formed (Fig. 6). We focus only on the major mechanisms of crustal transformations between different tectonic types, while other processes may exist as well.

5.1.1. Volcanic arcs

Volcanic arcs (we have only few locations of this type, see Table S1 for details) have a strongly heterogeneous crustal structure, which typically includes <12% of sediments, 30–50% of the upper-middle crust, and 50–65% of the mafic crust (Fig. 3). The northern part of the Honshu island is anomalous (Fig. 2). Its crust with 75% of the upper-middle crust and only ca. 23% of the mafic crust plots within the continental orogenic type, despite it is only 35 km thick beneath the Honshu island (Fig. 3b). Our results suggest the presence of a sizable continental fragment at Honsu. When the anomalous crust of the Honshu island is excluded, volcanic arcs have, on average, ca. 5% of sediments, 46% of the upper-middle crust, and 50% of the mafic crust (Fig. 2b) (Table 1).

5.1.2. Stable continents

Cratonic crust of shields and platforms, which covers huge areas in Eurasia, shows an extremely consistent pattern despite significant variations in the crustal thickness (Fig. 2). Cratonic crust differs from island arcs by a significant increase in the proportion of the upper-middle crust to 55–70% at the expense of the mafic crustal layer, which reduces to ca. 30–45% (Fig. 3a). This trend apparently reflects gradual formation of the felsic crust in island arcs (Gorton and Schandl, 2000), which includes underplat-

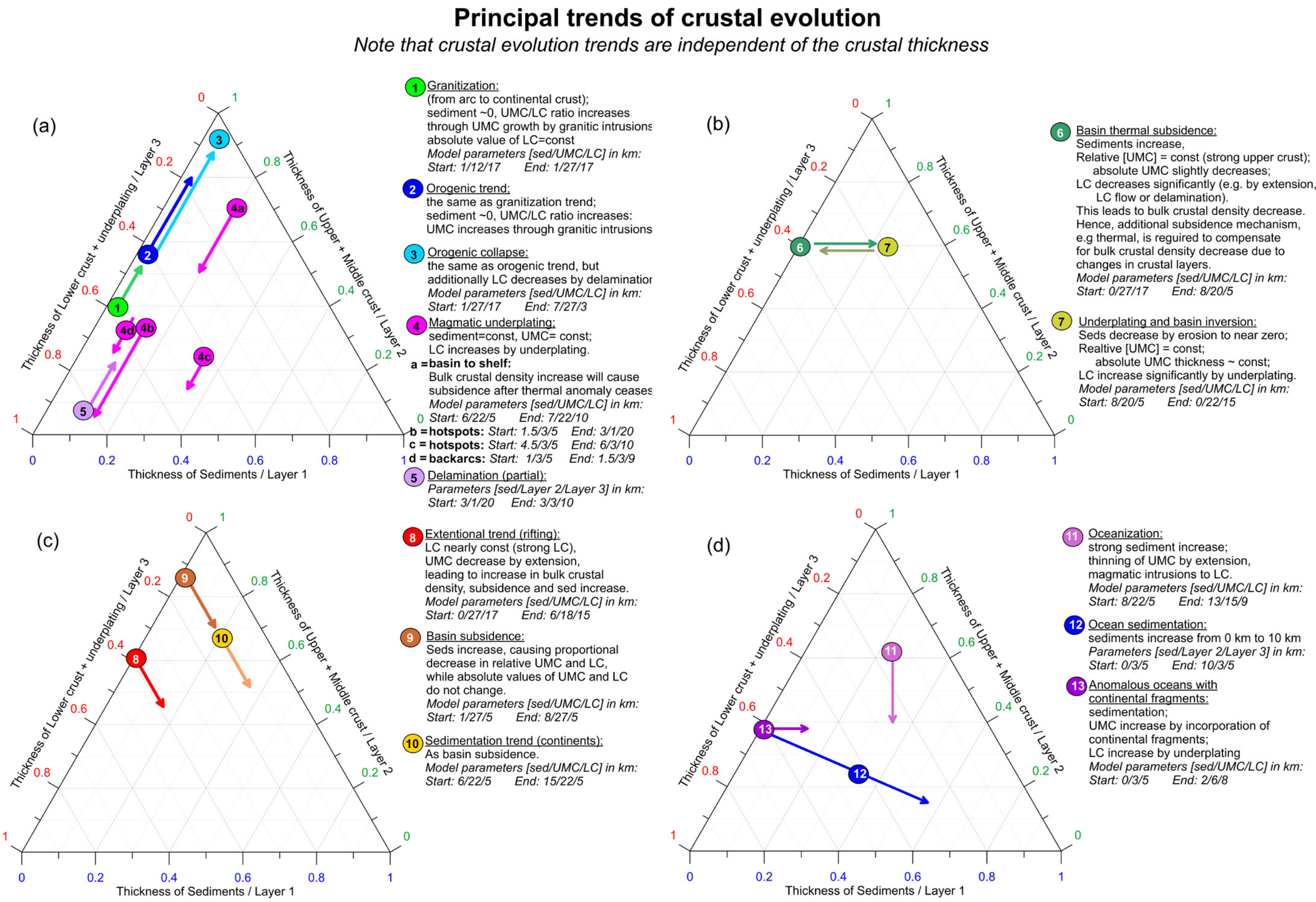

Fig. 5. Principal trends of crustal evolution: (a, b, c) when preserving relative thickness of one of crustal layers unchanged, (d) some trends for oceans. Abbreviations: sed = sedimentary layer; UMC = upper-middle crust (or Layer 2); LC = lower and high-Vp lowermost crust (or Layer 3).

ing of preexisting terranes by newly formed basaltic arc magmas, their interaction with pre-existing crust, differentiation and fractional crystallization with formation of felsic plutons which rise into the upper crust, increasing its thickness, while high-density mafic restites and cumulates eventually sink into the mantle (Lee et al., 2007) (Fig. 6, trend "G"). The observed trend does not seem to support a recent hypothesis that relamination, but not delamination of the densest parts of arc lower crust determines the average composition of continental crust (Kelemen and Behn, 2016). Further evolution of cratonic crust may go in three principal directions (Fig. 6): (i) orogenic trend (labelled "C" and "D1" in Fig. 6), (ii) basin trend (basin subsidence and sedimentation, labelled "S1" and "S2" in Fig. 6), which does not include changes in the basement layers and therefore is dominated by thermal and compositional subsidence (Kaminsky and Jaupart, 2000), and (iii) lower crustal delamination and sedimentation (labelled "S3" in Fig. 6).

5.1.3. Orogens

The orogenic trend ("C" and "D1" in Fig. 6) includes further granitization of the crust leading to an increase in the proportion of the upper-middle crustal layer. This happens by further growth of the felsic layer with the proportional decrease in the relative thickness of the mafic layer. The process may involve either further formation of granitic plutons above subduction systems associated with collisional orogens (Lee et al., 2015), or delamination of the lower crust (Kay and Kay, 1993).

There is little difference between Cenozoic and most of the Caledonian orogens, while the late Proterozoic-early Paleozoic massifs of the Variscan orogen form a clearly distinct group where

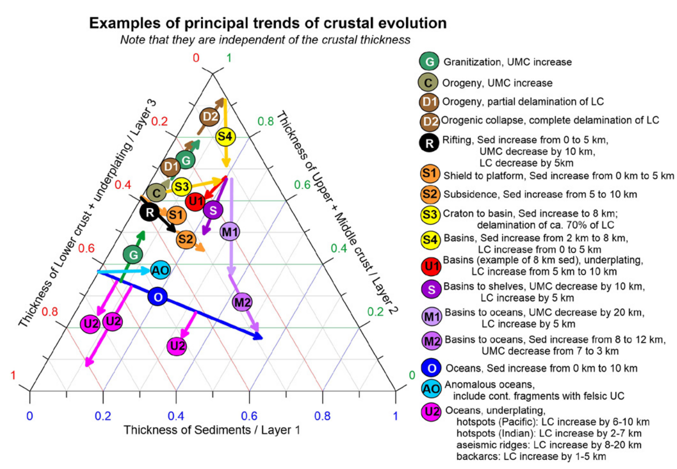

Fig. 6. Examples of some principal trends of crustal evolution, which cause a transformation of one crustal type to another. The starting/ending points of each process correspond to major crustal types shown in Fig. 2 (average crustal values for each tectonic class are used in calculations shown here). Some of the presented trends can be achieved through other transformations of crustal layers. Other transformation trends (not shown) are also possible. Abbreviations in legend: sed = sedimentary layer; UMC = upper-middle crust; LC = lower and high-Vp lowermost crust.

the mafic lower crust makes less than 10%, while the felsic-intermediate layer makes ca. 90% of the crustal thickness. Surprisingly, the Alps belong to the same group (Fig. 2a). We consider the Variscan orogens that have lost most of their lower crust (Aichroth et al., 1992; Artemieva and Thybo, 2013) as the end-member case for the orogenic trend, which includes lower crustal delamination ("D2" in Fig. 6) and orogenic collapse (Menard and Molnar, 1988) (trend "3" in Fig. 5). Further evolution of these collapsed orogens includes subsidence due to loss of crustal roots with consequent formation of sedimentary basins (trend "S4" in Fig. 6).

Crustal structure of several Paleozoic orogens is undistinguishable from cratonic platforms. These include the Norwegian and British–Irish Caledonides and the Ural mountains (Fig. 2). The Norwegian Caledonides is an allochtonous terrane atop the Precambrian lithosphere (Roberts, 2003), and our results suggest that a significant part of the British–Irish Caledonides (part of type O2 in Fig. 2 that plots with type C1–C3 in green oval) may be alike, while the unique crustal structure of the Urals (Brown et al., 2002) is commonly explained by preservation of this Paleozoic orogen in the intraplate setting where it was isolated from active tectonic deformation. Our results suggest that preservation of the Urals may be, at least in part, caused by its clearly craton-type crustal structure.

5.1.4. Rifts

The rifting trend of platform modification includes significant thinning of both the upper-middle and the lower crust accompanied by sedimentation (trend "8" in Fig. 5 and "R" in Fig. 6). The required thinning of basement layers (in relative values) is twice larger for the felsic layer than for the mafic one (10% and 5%, respectively). This suggests that either uniform crustal stretching as expected in the pure shear model (McKenzie, 1978) is not a common process, or that magmatic underplating and mafic intrusions into the lower crust compensate crustal thinning caused by extension, as proposed for the Baikal rift based on seismic interpretations (Thybo and Nielsen, 2009).

Several anomalous rifted structures are worth mentioning. The entire huge West Siberian Basin that has formed atop several large-scale Mesozoic rifts has crustal characteristics typical of rifts (Fig. 2). The Paleozoic Dnieper–Donets rift plots with other rifts only when the lower (high Vp and probably metamorphosed) part of a 20 km thick sedimentary sequence is considered as part of the basement (labeled DD, type E3 in Fig. 2a). However, when both young and paleosediments are considered as part of sedimentary layer, the crustal structure of the Dnieper–Donets rift appears truly unique and plots close to the ocean trend for normal oceans with a thick sedimentary cover (Fig. 2a). Our results therefore suggest that the Devonian Dnieper–Donets rifting may have initiated a continental break-up, but the process failed.

5.1.5. Basins

The results show that basins are best formed either by craton-to-platform transformation and by an orogenic collapse (trends "6" and "9" in Fig. 5). A subsidence trend leading to basin formation (the craton-to-platform trend labeled "S3" in Fig. 6) requires a 20% decrease of the mafic layer which leads to a 10% increase in the relative thickness of the felsic-intermediate layer and a 10% increase in sedimentation, if the starting point is cratonic crust (Fig. 2, green oval). All these processes reduce the average crustal density caused by compositional heterogeneity, and therefore it is not clear how they may cause subsidence rather than uplift (Fig. 5, trend "6"). Unless the total crustal thickness reduces, therefore reducing the total crustal buoyancy, this apparent controversy requires that the mantle plays an active role in basin subsidence (Kaminsky and Jaupart, 2000), such that lithosphere heating and subsequent cooling, enhanced by metamorphic reactions, should be a major factor. A plausible mechanism to put all these requirements in place would be a hot-spot impact on stable lithosphere, causing crustal heating by magmatic intrusions, followed by eclogitization and partial delamination of the lower crust, and thermal subsidence, accelerated by sedimentation.

Our results indicate a significantly different crustal structure of young Cenozoic basins (type B1) and old Phanerozoic basins (type B2), which include the Peri-Caspian Depression and Pechora Basin (labeled CN, CN*, and Pb in Fig. 2a). A sharp contrast between young and old basins may reflect a gradual progression in crustal metamorphism. Interestingly, on average, Meso-Paleozoic basins (type B2) plot together with the Arctic shelves (types S1/2 in Fig. 2b), therefore indicating the similarity between geodynamic processes responsible for their subsidence. Note that both the Peri-Caspian Depression and the Barents Sea shelf formed on the cratonic lithosphere.

5.1.6. Shelves

Evolution of basins may progress to formation of shelves as in the Arctic Ocean (Fig. 2b and trend "S" in Fig. 6). This process suggests significant crustal extension, thinning of the felsic-intermediate crustal layer, and possibly crustal mafication through magmatic intrusions with thickening of the mafic layer. Similarly, shelves may also be formed by subsidence of strongly rifted continental crust as has been proposed for the Barents Sea (Faleide et al., 1993) and the Bohai Basin in China (Hu et al., 2001) (trend "S2" in Fig. 6). The diversity of settings and processes in shelf formation is clear from Fig. 2a, where they extend from basins (like the Western Barents Sea) to rifts (like the East and West Kara Sea, where the rift system of the West Siberian Basin continues off-shore). It is interesting that crustal structure of the Baffin Bay and the Labrador Sea (labeled "failed oceans" because the North Atlantic spreading failed there and jumped to the eastern side of Greenland) is very similar to the Arctic shelves (Fig. 2a), further confirming that rifting stage is very important in shelf evolution.

5.1.7. Failed oceans

Failed oceans include the Baffin Bay, the Labrador Sea, northernmost part of the North Atlantic Ocean and, very surprisingly, the Peri-Caspian Basin and the Black Sea. Our analysis does not support the hypothesis for the back-arc origin of the Black Sea (labeled M4* in Fig. 2b) (Zonenshain and Pichon, 1986), but rather indicates that its formation was similar to the Peri-Caspian Basin (Fig. 2a), where the presence of the Devonian rifts is recognized in the crustal structure (Nikishin et al., 1996). The anomalous crust of the Peri-Caspian Basin may reflect late stages of rifting (trend "11" in Fig. 5), with thinning of the felsic layer and thickening of the mafic layer ("M1" in Fig. 6); further evolution may lead to ocean formation ("M2" in Fig. 6). However, our results do not confirm the presence of oceanic crust in the Peri-Caspian Basin (Fig. 2).

5.1.8. Anomalous oceans

Normal ocean is not present in the North Atlantic region, where all provinces plot above the oceanic sedimentation trend ("O" in Fig. 6). Anomalous oceanic crust of the type observed in the North Atlantic region around Iceland (Fig. 2) cannot be formed only by magmatic intrusions, as may be expected in ocean regions with hot-spots ("U2" in Fig. 6). Our results for the anomalous parts of the North Atlantics suggest that during ocean opening oceanic crust may have incorporated some continental fragments with felsic material ("AO" in Fig. 6), such as expected at the Jan Mayen block (Peron-Pinvidic et al., 2012). The presence of a strong chemical heterogeneity in the North Atlantic region is in agreement with regional geochemical studies of oceanic peridotites (Korenaga and Kelemen, 2000).

5.1.9. Back-arc basins

Our analysis includes only 2 back-arc basins, the Sea of Japan and the Lau Basin (Iwasaki et al., 2013; Crawford et al., 2003). They plot below the oceanic trend (Fig. 2), because the lower crustal layer (or oceanic Layer 3) constitutes 65–80% of the crust. Back-arc basins have a large variability in the crustal structure, both within the individual back-arcs and between them (Iwasaki et al., 2013), however limited data do not allow for further speculations.

5.1.10. Oceanic hotspots and aseismic ridges

The crustal structure of the Pacific hotspots is truly unique, and these are the only tectonic structures where the lower crustal layer (oceanic Layer 3) constitutes ca. 90% of the crust (Fig. 2). Such structure may be produced by magmatic additions (underplating) to normal oceanic crust (“U2” in Fig. 6). The same trend indicates that the Indian ocean hotspots can be formed by the same process, but in oceans with a large sequence of sediments (trends “4b” and “4c” in Fig. 5). Further evolution of the hotspot crust should include delamination of underplated material (especially if metamorphosed to a higher density mineral phase), which would shift the crust of the Pacific hotspots towards aseismic ridges (trend “5” in Fig. 5).

Aseismic ridges plot between normal oceans with ca. 1.5 km of sediments and the Pacific hotspots. The absolute values of the crustal thickness suggest that they formed not from hotspots (Fig. 3b and Table S1), but from normal oceans the same way as the hotspots. Surprisingly, aseismic ridges overlap with the back-arc basins of the western Pacific (Fig. 2), suggesting that the crust of the Sea of Japan and the Lau Basin has also been modified by magmatic additions, similar to hotspot tracks.

5.2. Crustal density

Knowledge of bulk crustal density is critical for different types of geodynamic modeling, including gravity modeling and buoyancy analysis, however the current state of knowledge remains unsatisfactory. Most approaches exploit correlation between seismic velocities and density derived from a combination of seismic, drilling, and laboratory measurements (Carlson and Henrick, 1990; Christensen and Mooney, 1995). This approach leads to a large variability of reported values due to non-uniqueness of velocity-density conversion. It is commonly assumed that typical oceanic crust has a density of ca. 2900 kg/m^3 (Carlson and Riskin, 1984), with reported values ranging from 2800 to 3000 kg/m^3, and the density of continental crust is around 2700–2800 kg/m^3, also with a significant scatter in reported values.

Clearly, crustal composition controls average crustal density. Our analysis allows for calculating average crustal density depending on specific tectonic setting. Fig. 2 shows that each crustal type is characterized by a very narrow range of relative thicknesses of the three major crustal layers. By assuming typical densities for crustal layers (2500 kg/m^3 for sediments (L1), 2750 kg/m^3 for the upper-middle crust (L2), and 3000 kg/m^3 for the lower and high-Vp lowermost crust (L3)), we calculate typical density of bulk crust in different tectonic settings (Fig. 7). We chose the same density values for three principal crustal layers, no matter the tectonic setting, to allow for a global comparison between transitional crust (failed oceans, anomalous oceans, back-arcs, island arcs, and highly extended continental crust of basins, rifts, and shelves), oceanic crust, and continental crust. Our choice of the density values is supported by laboratory data and petrological studies of typical continental and oceanic crust (Carlson and Herrick, 1990; Christensen and Mooney, 1995; cf. Rudnick and Fountain, 1995).

The results show that the average crustal density is systematically lower in continental crust than in oceanic crust, except for some anomalous oceanic settings with a large thickness of sediments. On continents, the average crustal density varies from 2700–2750 kg/m^3 in sedimentary basins, to 2750–2800 kg/m^3 in orogens, rifts, and shelves, to 2800–2850 kg/m^3 in shields and platforms, reaching the highest values of 2850–2875 kg/m^3 in island arcs. In oceans, bulk crustal density decreases from

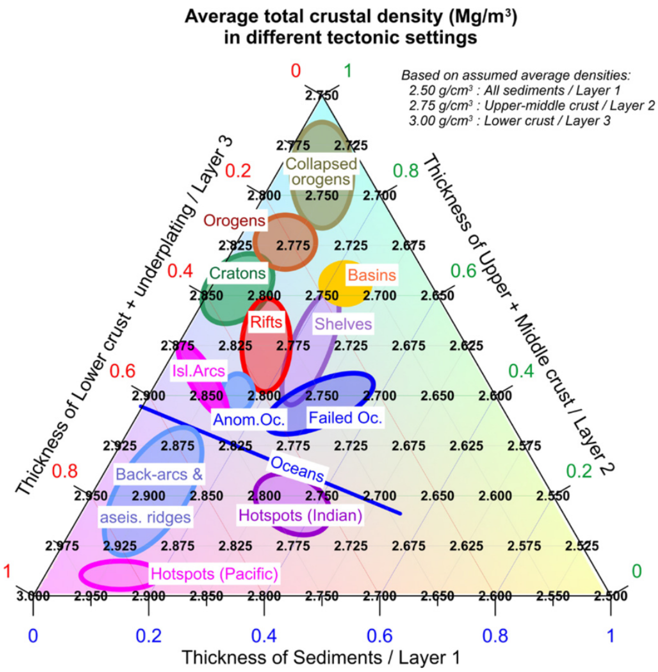

Fig. 7. Typical values of average density of oceanic and continental crust in different tectonic settings. Color code for crustal provinces is the same as in Figs. 1, 2, 4. Bulk crustal densities are calculated by assuming typical densities for three crustal layers: 2500 kg/m^3 for sediments/Layer 1; 2750 kg/m^3 for the upper-middle crust/Layer 2, and 3000 kg/m^3 for the lower and high-Vp lowermost crust/Layer 3.

2900 kg/m^3 in normal oceanic crust without sediments to 2830 kg/m^3 and 2700 kg/m^3 when sediments reach 3 km and 10 km in thickness, correspondingly. Back-arc basins in oceanic settings and aseismic ocean ridges have an average crustal density of ca. 2875–2925 kg/m^3, and globally the highest density values of 2900–2950 kg/m^3 are typical of the Pacific Ocean hotspots, while in contrast, the Indian Ocean volcanic islands have a low density crust (2750–2800 kg/m^3).

6. Conclusions

We propose a new method of crustal typization based on the analysis of the relative thicknesses of three major crustal layers.

1. We demonstrate that the ratio of these relative thicknesses uniquely characterizes different crustal types. This crustal characteristic is fundamental and is completely independent of the total crustal thickness, which therefore does not appear to be an important parameter to typify the crust globally.

2. The new method allows for a simple description of major trends of crustal evolution, modification and transformation and for recognizing genetic links between the crust of different geodynamic origin. We discuss these geodynamic trends and illustrate them by regional examples. In particular, our results demonstrate an importance of intracratonic rifting in shelf formation. We also show that, by relative thickness of crustal layers, several Paleozoic orogens (the Norwegian and British–Irish Caledonides and the Ural mountains) are undistinguishable from cratonic platforms.

3. The new method allows for recognizing the origin of the crust in regions with poorly known geology, such as in ice-covered regions of Greenland and Antarctica, and in submarine regions like the Arctic shelf. In particular, our results indicate the presence of continental fragments at the Japan volcanic arc beneath the Honsu island and do not support the hypothesis of a back-arc origin of the Black Sea, where normalized crustal structure is similar to the Peri-Caspian Basin and to the North Atlantic domains with failed ocean spreading.

4. As a practical application, we calculate average crustal density in different tectonic settings. The results show a significant variability in average density of continental crust, which has lower density than oceanic crust in most tectonic settings.

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Appendix A. Supplementary material

Supplementary material related to this article can be found online at https://doi.org/10.1016/j.epsl.2019.01.033.

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