Continental subduction and a mechanism for exhumation of high-pressure metamorphic rocks: new modelling and field data from Oman

Alexander I. Chemenda a,*, Maurice Mattauer a, Alexander N. Bokun b

a Laboratoire de Géophysique et Tectonique. UMR 5573. Université Montpellier II. Case 060, Pl. E. Bataillon-34095, Montpellier, Cedex 05, France
b Institute of Geology and Geochemistry of Fuel Minerals, Lvov, Ukraine

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Abstract

The physical model presented reveals two principal regimes of continental subduction: a highly compressional (HC) regime and a low compressional (LC) regime characterised by high and low pressure between the overriding and subducting plates, respectively. The pressure is inversely proportional to the pull force, which depends on the difference between average density of the subducting lithosphere and density of the mantle. The subducting continental crust reaches a maximum depth which is proportional to the strength of the crust, inversely proportional to the interplate pressure, and is ca. 200 km on average. The crust then fails, forming a major crustal thrust. The location of failure depends on the regime of subduction (on the interplate pressure): in the HC regime, failure occurs near the front of the subduction (collision) zone, whereas the crust fails at greater depth under the base of the overriding plate in the LC regime. The failure is followed by buoyancy-driven uplift of the subducted crustal slice, while the lithospheric mantle keeps subducting. The uplift causes a normal sense displacement (formation of a normal fault) along the upper surface of the crustal slice. For the LC regime (failure of the crust near the base of the overriding plate), a rapid spontaneous crustal uplift (intrusion into the interplate zone) brings the deeply subducted crust to shallow depths. Under the pressure of this crust the frontal part of the overriding plate undergoes local extension and then fails, forming a tectonic window. The rising material (the high-pressure rocks) is exhumed within this window. In the HC regime, uplift of the subducted crust, after its failure in front of the overriding plate, is possible only with erosion of the relief, which provides an unloading effect, allowing the subducted crustal slice to rise up. The exhumation depth is generally smaller in this regime but the volume of the exhumed material is larger. The HC subduction regime has been shown earlier to match the Himalayan situation. The LC regime fits the situation in the Oman Mountains considered in this paper.

Keywords: subduction; orogeny; exhumation; metamorphic rocks; models; Oman Mountains

* Corresponding author. From October 1, 1996 at: UMR GeoSciences Azur, Université de Nice Sophia Antipolis, 250 Rue Albert Einstein-Sophia Antipolis, 06560 Valbonne, France. Tel.: +33 67144690. Fax: +33 67523908. E-mail: shem@dstu.univ-montp2.fr

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1. Introduction

This work follows the physical modelling of continental subduction developed by Chemenda et al. [1,2] within the overall framework shown in Fig. 1. The main results obtained in previous experiments were the following: At the initial stages, subduction of the continental lithosphere develops similarly to the preceding subduction of the oceanic lithosphere (Fig. 2a). The continental crust subducts, together with the lithospheric mantle, to a critical depth of 150 km or more, which has been shown to be strongly dependent on and proportional to the effective strength (yield limit) of the upper crustal layer, \( \sigma_{\text{uc}} \). Upon reaching this depth, the upper crustal layer fails in front of the subduction zone (Fig. 2b). The failure is followed by the formation of a mountain belt, due to underthrusting and thickening of the crust. Erosion (mechanical removal in the model) of part of the forming relief produces an unloading effect, causing buoyancy-driven uplift of the subducted crustal slice (Fig. 2c,d). This uplift results in both exhumation of the subducted crustal material and formation of a major (10–100 km displacement) normal fault along the upper surface of the ascending slice (Fig. 2c,d). The exhumation depth depends on the depth of the preceding crustal subduction, or on the effective strength \( \sigma_{\text{uc}} \) of the upper crust: the larger this depth (the \( \sigma_{\text{uc}} \) value), the greater the buoyancy-driven uplift of the crustal slice. This scenario fits the situation in the Himalayas [1].

In this work we define another physically possible regime of continental subduction. We found that, besides \( \sigma_{\text{uc}} \), there is a second parameter controlling both the process of the crustal subduction and the following exhumation. This parameter is the inter-
Fig. 3. Forces in the subduction zone. The full interplate pressure $P_n$, the pressure force $F_P$, and the horizontal compression of the lithosphere (horizontal component $F_h = \int_0^H P(z) \, dz$ of the pressure force) decrease with increasing pull-force, $F_F$ and/or the super hydrostatic pressure $\Delta P$ in the mantle. $\Delta P$ acts on the surface of the subducted plate due to the relative horizontal motion of this plate with respect to the surrounding mantle (see [2] for details). $F_b$ is the along-plate component of the buoyancy force.

plate pressure, $P_n$ (Fig. 3), which depends largely on the density contrast $\Delta \rho = \rho_o - \rho_a$ ($\rho_o$ is the average density of the subducting lithosphere and $\rho_a$ is the asthenospheric density) or, more exactly, on the pull force, which is proportional to $\Delta \rho$. The lower this contrast (the higher the pull force), the lower the pressure $P_n$ between the plates [2]. Reduction of the pressure increases the depth of crustal subduction but also changes the location (depth) of crustal failure: with low pressure the crust fails under the base of the overriding plate. Deep crustal failure changes both style and depth of the following exhumation, leading to the appearance on the surface of high to ultrahigh pressure rocks. This process has natural analogues, the Oman Mountains seeming to be one of them. The geological evolution of this region fits well with the model.

2. Experimental setting

The experimental setting (Fig. 1), model materials, techniques, and similarity criteria are the same as those used in our previous work [1]. The difference between the present model and previous trials is that in this experiment we have varied (increased) the average density of the subducting continental lithosphere (and hence the pull force) by either reducing the thickness of the crust, or increasing the density of the mantle layer, or both.

3. Results of the experiments

In [1] we conducted experiments with the parameter values presented in Table 1. According to the similarity criteria these values correspond to the following values for prototype [1]: $\sigma_1 \approx \sigma_2 = 5.7 \times 10^8$ Pa; $\sigma_1 \approx 1.1 \times 10^7$ Pa; $\rho_1 = 3.4 \times 10^3$ kg/m$^3$; $\rho_c = (2.8-2.9) \times 10^3$ kg/m$^3$; $\rho_a = 3.3 \times 10^3$ kg/m$^3$; $H_f = 6 \times 10^4$ m; $H_c = 3.2 \times 10^4$ m; $H_c = 0.8 \times 10^4$ m; $V = 5$ cm/yr (see Table 1 for notations). The time ratio is 1 min in the model to 0.5 Ma in nature. In these experiments the continental crust subducted to a depth of about 5 cm (corresponding to 200 km in nature) before crustal failure

Table 1

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Experiment in [1]</th>
<th>Experiment 1</th>
<th>Experiment 2</th>
<th>Experiment 3</th>
</tr>
</thead>
<tbody>
<tr>
<td>$a_1$ (Pa)</td>
<td>43</td>
<td>43</td>
<td>43</td>
<td>Parameter values are the same as in exp. 2, except that the length of the frontal segment of the subducting plate without crust is increased by 2 mm</td>
</tr>
<tr>
<td>$a_2$ (Pa)</td>
<td>43</td>
<td>43</td>
<td>43</td>
<td>Parameter values are the same as in exp. 2, except that the length of the frontal segment of the subducting plate without crust is increased by 2 mm</td>
</tr>
<tr>
<td>$a_3$ (Pa)</td>
<td>0.8</td>
<td>0.8</td>
<td>0.8</td>
<td>Parameter values are the same as in exp. 2, except that the length of the frontal segment of the subducting plate without crust is increased by 2 mm</td>
</tr>
<tr>
<td>$\rho_f$ (g/cm$^3$)</td>
<td>1.03</td>
<td>1.03</td>
<td>1.03</td>
<td>Parameter values are the same as in exp. 2, except that the length of the frontal segment of the subducting plate without crust is increased by 2 mm</td>
</tr>
<tr>
<td>$\rho_c$ (g/cm$^3$)</td>
<td>0.86</td>
<td>0.86</td>
<td>0.86</td>
<td>Parameter values are the same as in exp. 2, except that the length of the frontal segment of the subducting plate without crust is increased by 2 mm</td>
</tr>
<tr>
<td>$\rho_a$ (g/cm$^3$)</td>
<td>1.5</td>
<td>0.8</td>
<td>0.2</td>
<td>Parameter values are the same as in exp. 2, except that the length of the frontal segment of the subducting plate without crust is increased by 2 mm</td>
</tr>
<tr>
<td>$H_f$ (cm)</td>
<td>0.5</td>
<td>0.8</td>
<td>0.2</td>
<td>Parameter values are the same as in exp. 2, except that the length of the frontal segment of the subducting plate without crust is increased by 2 mm</td>
</tr>
<tr>
<td>$H_c$ (cm)</td>
<td>1.5</td>
<td>1.5</td>
<td>0.2</td>
<td>Parameter values are the same as in exp. 2, except that the length of the frontal segment of the subducting plate without crust is increased by 2 mm</td>
</tr>
<tr>
<td>$V$ (m/s)</td>
<td>10$^{-4}$</td>
<td>10$^{-4}$</td>
<td>10$^{-4}$</td>
<td>Parameter values are the same as in exp. 2, except that the length of the frontal segment of the subducting plate without crust is increased by 2 mm</td>
</tr>
</tbody>
</table>

$\sigma_1$, $\sigma_2$, and $\sigma_3$ are the yield limits for the mantle and upper and lower crustal layers of the lithosphere under normal load, respectively; $\rho_f$, $\rho_c$, $\rho_a$, $\rho_a$, $\rho_c$, $\rho_f$ are the densities of the mantle lithospheric layer, asthenosphere, upper and lower crustal layers, respectively; $H_f$, $H_c$, and $H_c$ are the thicknesses of the mantle lithospheric layer, upper and lower crustal layers; $V$ is the rate of the plate convergence.
in front of the subduction zone (see Fig. 2b). Below we present a new experiment conducted under similar conditions, but with the thickness of the upper crustal layer reduced by 2 mm (the strength of the overriding plate in all experiments is equal to that of the upper crust; the density of this plate is equal to the asthenospheric density).

Experiment 1 (Fig. 4): In this experiment the crust subducts to a depth of near 7 cm (300 km in nature) versus 5 cm (200 km) in the previous experiments. Further continuation of the experiment was impossible because the mantle layer of the subducted plate reached the bottom of the bath. A continuation of subduction would result in failure of the crust due to...
the increasing buoyancy force $F_b$ (see Fig. 3) which is proportional to the depth of crustal subduction. To obtain this failure with the same experimental setting (with the same height of the bath) one must reduce the strength of the upper crust, which was done in the next experiment.

**Experiment 2** (Fig. 5): In this experiment the yield limit of the upper crustal layer $\sigma_i^m$ is reduced by two times to $\sigma_i^m = 22$ Pa (which corresponds to 300 MPa in nature); $\rho_i^m = 1.04 \times 10^3$ kg/m$^3$ (superscript $m$ denotes model parameters). Other parameters are the same as in the previous experiment (Table 1). In this experiment, the crust subducts to a depth near 4 cm (about the same as in the experiments illustrated in Fig. 2 with two times stronger upper crust). The crust then fails at depth under the base of the overriding plate (Fig. 5c). The subducted crustal slice starts to uplift, tending to intrude the interplate zone but without being able to separate the plates. The crustal material accumulates at the base of the overriding plate, producing local elevation and extension of this plate (the extension was observed on the surface over the accumulated crust).

To facilitate separation of the overriding and subducting plates and allow intrusion of the ascending crustal slice into the interplate zone, we increased the pull force in the next experiment.

**Experiment 3** (Fig. 6): We slightly increased (by 2 mm) the length (in cross section) of the ‘continental margin’ (the length of the leading segment of the subducting plate without crust). This small modification allowed the subducted crustal slice to intrude after failure into the interplate zone and to rise between the plates (Fig. 6a). The rising slice produces elevation and extension of the overriding wedge of the upper plate, resulting in break-up of its frontal, thinnest segment (Fig. 6b). We did not apply ‘erosion’ in this experiment, although it is clear that erosion of the local elevation of the overriding plate would have facilitated break-up (lifting) of the overriding plate and hence uplift of the crustal slice.

4. Analysis of the experimental results and discussion

The difference between the new regime (mode) of continental subduction shown in Fig. 7 and the previous model in Fig. 2 is caused by the different interplate pressure $P_n$ exerted by the subducting plate on the overriding plate. The pressure is controlled by the pull force $F_p$ (Fig. 3) which is itself inversely proportional to the $\Delta p = \rho_a - \rho_s$ value ($\rho_a$ is the average density of the subducted lithosphere and $\rho_s$ is the asthenospheric density). In reality, $P_n$ also depends on other factors, such as the pull force from previously subducted oceanic lithosphere and the relative horizontal motion between the lithosphere and the underlying mantle (Fig. 3) [2,3]. A high pressure $P_n$ causes high compressive stresses within the crust in front of the subduction zone (both pressure and buoyancy force act against crustal subduction). This results in failure of the crust in front of the subduction zone (Fig. 2b) and in a strong compression (high force, $F_b$ in Fig. 3) of the overriding lithosphere (highly compressional subduction regime). A lower pressure decreases compression of the crust and, hence, of the whole lithosphere (low compressional subduction regime), which allows the crust to subduct considerably deeper without failure. On the other hand, reduction of the compressive stresses in front of the subduction zone shifts the location of crustal failure (which can now be due to buoyancy force only) to greater depth, near the base.
of the overriding plate. Low pressure between the plates then allows the buoyant, subducted crust to intrude the interplate zone (Fig. 7b).

The efficiency of a low compressional (LC) mode of continental subduction (Fig. 7) in exhuming subducted material is substantially greater than of a

Fig. 8. Evolutionary model of Oman (stages (b), (c) and (d) correspond to the period from 90–80 to 80–70 Ma). (a) Geological profile through Oman to the east of Muscat (the arrows correspond to the stage of exhumation). (b) Failure of the crust near the base of the overriding plate (ca. 60 km depth). (c) Rapid, buoyancy-driven uplift of the subducted crustal slice which scrapes and pushes up the sediments dragged down into the interplate zone during previous stages of the oceanic and continental subduction. (d) Breakage of the overriding wedge and formation of the window with exhumed sediments. (e) End of rapid tectonic uplift of the subducted crust and break-off of the dense, subducted mantle layer, with part of the lower crust metamorphosed to a density in excess of the asthenospheric density. (c) Present situation. 1 = Arabian upper (a) and lower (b) crust; 2 = continental sedimentary cover (a = Permian-Mesozoic; b = Proterozoic and Paleozoic); 3 = oceanic lithosphere; 4 = oceanic sediments (Hawasina nappe); 5 = ductile fault; 6 = cleavage and folds; 7 = supposed present geometry of the lithospheric base; 8 = thrust (a) and normal (b) faults; 9 = marker corresponding to 60 km depth (ca. 20 kbar) at stage in (b); 10 = erosion; 11 = direction of regional vertical movement after the break-off of the mantle layer.
highly compressional (HC) regime (Fig. 2) because it provides considerably deeper crustal subduction at the same strength of the upper crust and does not require erosion (the exhumation in the HC regime in Fig. 2 does not occur without erosion). When the subducted crustal slice intruding the interplate zone reaches the surface, the exhumed material has already been uplifted from a great depth (Fig. 7). Further uplift (exhumation) can be continued by the same ‘erosion activated’ mechanism as in the HC mode in Fig. 2 until the reducing (due to the uplift of the crust) buoyancy force is balanced by the forces acting against uplift [1].

There is another possible scenario of exhumation corresponding to the LC regime. The uplifting crustal slice does not necessarily reach the surface, producing rifting within the overriding plate, as shown in Figs. 6 and 7. The rifting does not occur in the experiments with stronger overriding plate. The initially rapidly uplifting crust stops in these experiments at some shallow depth, corresponding to a few tens of kilometres in nature. After that, further exhumation can be slowly produced by erosion.

As already mentioned, one of the major factors defining the subduction mode is the density contrast \( \Delta \rho \). In the experiments corresponding to a low compressional mode (experiments 2 and 3), \( \Delta \rho^m = 0.02 \times 10^3 \text{ kg/m}^3 \), which corresponds to \( \Delta \rho^o = 0.065 \times 10^3 \text{ kg/m}^3 \) in nature (superscripts m and o denote model and original (nature) parameters). According to Cloos [4], such a density contrast can characterize a 100 km thick continental lithosphere including a 26 km thick granitic \( (\rho_c = 2.75 \times 10^3 \text{ kg/m}^3) \) crust. For a highly compressional mode (Fig. 2), \( \Delta \rho^m = 0.038 \times 10^3 \text{ kg/m}^3 \) and \( \Delta \rho^o = 0.124 \times 10^3 \text{ kg/m}^3 \) [1]. To have this density contrast for a 100 km thick lithosphere, the thickness of the granitic crust should be at least 35 km [4]. Of course, one can obtain the same \( \Delta \rho^o \) value by assuming a different composition (density) and/or thicknesses of the crustal and mantle lithospheric layers. There is one additional factor which strongly affects \( \Delta \rho^o \): the mineralogical transformations of the subducting crust can increase its density to \( 3 \times 10^3 \text{ kg/m}^3 \) or even more [5]. This factor depends on the composition of the crust, the presence of water, the kinetics of the metamorphic reactions, and the thermal regime of the subducting crust. These are poorly constrained and cannot be properly incorporated into a simple model. Whatever the values and combinations of the parameters defining the interplate pressure, the crust fails upon reaching a critical (for a given set of parameter values) depth. There are only two possible failure locations, either near the front of the overriding plate (HC regime) or near the base of this plate (LC regime). Whether these regimes actually take place in nature, and which one occurs in a given region, cannot be answered solely from physical constraints because of the mentioned uncertainties. On the other hand, both regimes have clear and different geological consequences which can be tested against geological data. It has been shown by Chemenda et al. [1] that the geological situation in the Himalayas corresponds to a high compressional subduction regime shown in Fig. 2. It seems that the regime presented in Fig. 7 also has natural analogues. One of them is the Oman mountain belt.

5. Model of continental subduction in Oman

The Oman Mountains are famous and well studied, mainly because of their spectacular ophiolite nappe (Fig. 8a) [6]. The nappe represents a frontal segment of oceanic lithosphere overthrust (obducted) on the Arabian continent during the upper Cretaceous [7]. The sediments of the Arabian Plate dragged under the nappe are exposed at present within the Saih Hatat window [8]. These sediments underwent intensive plastic deformation with folding and metamorphism of blueschist to eclogite facies [9]. Both deformation and metamorphism increase towards the north, with the maximum degree of eclogitization corresponding to ca. 20 kbar and 500°C [10,11]. K–Ar and \(^{40}\)Ar/\(^{36}\)Ar ages for the metamorphism are 72–80 Ma [12,13]. The deformation of the sediments is very complex and polyphase [11,14,15]. The geological observations reveal, however, two major phases with parallel NNE striking stretching lineations [11,14]. The older phase corresponds to the subduction of the continental margin. It is marked by a northward dipping schistosity and isoclinal folds overturned to the south. This stage was followed by the opposite sense of shear during exhumation of the subducted material, with formation of northward dipping normal faults in the northern part of the Saih...
Hatat (Fig. 8a). The drastic metamorphic contrast (up to 0.6 GPa) documented along the normal faults [11] implies great displacement. In the southern part of the Saih Hatat window, the exhumation phase is marked by southward dipping thrusts and cleavage associated with folding overturned to the North [15] (Fig. 8a).

The Upper Cretaceous evolution of Oman is relatively well studied [6,16–18], although pre-abduction regional plate tectonics and the origin of the Oman (Semail) ophiolite are still poorly understood. What is clear is that the north-dipping intra-oceanic subduction located to the north of Oman was followed at near 90 Ma by abduction (i.e., by subduction of the Arabian continental margin under the overriding oceanic plate): 200 km [17] to 600 km [16] of the transient and continental lithosphere was subducted in this zone. The obduction stage was followed by splitting of the obducted oceanic wedge (separation of the Semail ophiolite klippe) and a rapid [19] exhumation of previously subducted sediments of the continental margin within the Saih Hatat window (Fig. 8a). The mechanism of this process is not well understood and still remains a subject of discussion [9,11,15,20,21]. Basically, the question is how normal faulting and associated rapid exhumation of HP rocks within the Saih Hatat anticline combine with coeval thrusting and regional plate convergence during the formation of the Oman belt [11]. Similar problems exist in other collisional belts, such as, for example, the Alps, the Urals, the Dabie Shan, and the Himalayas. Our model of continental subduction provides a possible physical basis for a consistent solution of this problem. The geological data presented suggest that continental subduction in Oman was characterised by a low compressional regime (Fig. 7). A possible corresponding evolutionary model is presented in Fig. 8. According to the model, the crust of the Arabian margin subducts to a depth of about 200 km before failure at a depth of ca. 60 km (Fig. 8b). The subducted crustal slice then intrudes the interplate zone, scrapes, and pushes up the weak heated sediments (both oceanic and continental, Fig. 8c) which were dragged down during previous stages of subduction. The frontal segment of the overriding plate is uplifted and undergoes local extension, due to both thickening of the crust (doubling of the upper crust) and overpressure in the sediments pushed from below by the rising crust. At the next stage (Fig. 8d), the thin frontal part of the overriding plate is broken up. This process is facilitated by erosion of the forming elevation. Both the deep sediments pushed upwards along the interplate zone and the sediments at higher levels (directly under the rupture zone) come to the surface and push the broken portion of the overriding plate (the Semail ophiolite klippe) towards Arabia (Fig. 8d). Such a scenario is consistent with the north-dipping ductile normal faults observed in the northern part of the Saih Hatat anticline and with the north-vergent folds and thrusts in the southern part of the anticline shown in Fig. 8a (see [22] for more detailed geological analysis of the model presented).

The removal (uplift) of the subducted crust from the lithospheric mantle layer increases gravitational instability of this layer (increases the pull force), which results in it breaking off (Fig. 8d). Removal of a dense mantle root reduces the pull force, which is part of the force driving subduction. The plate convergence can stop. If it continues under the external (push) forces, then there are two possibilities: The continental subduction in Oman can either continue, shifting to a HC regime, or the lithosphere can fail elsewhere under the increasing compression, resulting in a jump of the zone of lithospheric consumption. It seems that the latter possibility was realised in the region in question: the deformation in the zone of continental subduction in Oman ceased with the initiation of a new subduction zone in Makran.

6. Conclusions

This model reveals two possible regimes of continental subduction: a highly compressional (HC) regime, considered by Chemenda et al. [1], and a low compressional (LC) regime obtained in the experiments reported here. Similarly to oceanic subduction, which is characterized by two major regimes (extensional and compressional [2]), the regimes of continental subduction depend on the pull force; the HC regime corresponds to a low pull force and the LC regime to a high pull force. A high pull force can be
provided by: (1) a relatively thin subducting continental crust; (2) a very thick, dense lithospheric mantle; (3) eclogitization of the subducted crust (especially of its basaltic part); and (4) a pull force from previously subducted oceanic lithosphere. A high pull force, in turn, results in low interplate pressure, which, on the one hand, facilitates the subduction of the buoyant crust to great depth and, on the other hand, localizes the failure of the subducted crust under the base of the overriding plate. The depth of crustal subduction also strongly depends on and is proportional to its strength. Deep subduction of a strong crust reduces the average density of the subducted plate and hence the pull force. Therefore, the crustal strength also influences the subduction regime and the location of crustal failure. The probability to fail at depth under the interplate zone is higher for a weak crust. After the breakage, the subducted crustal segment rises, intrudes the interplate zone, and pushes up the sediments previously dragged down into this zone. Such a process produces local elevation, extension and break-up (riifting) of the frontal, thin and weak segment of the overriding plate. Another consequence of this process is the formation of a major normal fault corresponding to the surface of the rising crustal slice. Break-up of the overriding plate is followed by exhumation, first of the sediments and then of the upper part of the uplifting crustal slice itself (the rising crustal slice may stop at some depth and then be exhumed by erosion). These processes develop during continuing plate convergence.

This model was tested in a relatively simple mountain belt in Oman and was shown to fit the geological situation in this region where sediments of the continental margin were first subducted to a depth of about 60 km and then rapidly returned back to the surface. Preliminary analysis shows that the model can be applied to other collisional belts characterized by syn-collisional exhumation of HP and UHP rocks, such as the Alps, the Urals, the Dabie Shan, the Caledonian and the Variscan belts. At least some of these belts seem to have undergone more complex evolution including both LC and HC regimes of continental subduction. The regime depends on the pull force, which may change during subduction; due to, for example, the subducted mantle lithosphere breaking off.

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References


