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# Exhumation of high-pressure rocks of the Kokchetav massif: facts and models

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**Abstract**

The exhumation of ultrahigh-pressure (UHP) metamorphic units from depths more than 100–120 km is one of the most intriguing questions in modern petrology and geodynamics. We use the diamondiferous Kumdy-Kol domain in the Kokchetav Massif to show that exhumation models should take into consideration initially high uplift velocities (from 20 down to 6 cm/year) and the absence of the deformation of UHP assemblages. The high rate of exhumation are indicated by ion microprobe (SHRIMP) dating of zircons from diamondiferous rocks and supported by the low degree of nitrogen aggregation in metamorphic diamonds.

Diamondiferous rocks in the Kumdy-Kol domain occur as steeply dipping (60°–80°) thin slices (few hundred metres) within granite–gneiss. Using geological, petrological and isotopic–geochemical data, we show that partial melting of diamondiferous metamorphic rocks occurred; a very important factor which has not been taken into account in previous models.

Deformation of diamondiferous rocks at Kumdy-Kol is insignificant; diamond inclusions in garnet are often intergrown with mica crystals carrying no traces of deformation. All these facts could be explained by partial melting of metapelites and granitic rocks in the Kumdy-Kol domain. The presence of melt is responsible for an essential reduction of viscosity and a density difference ($\Delta\rho$) between crustal rocks and mantle material and reduced friction between the upwelling crustal block, the subducting and overriding plates. Besides $\Delta\rho$, the exhumation rate seems to depend on internal pressure in the subducting continental crustal block which can be regarded as a viscous layer between subducting continental lithosphere and surrounding mantle.

We construct different models for the three stages of exhumation: a model similar to "corner flow" for the first superfast exhumation stage, an intermediate stage of extension (most important from structural point of view) and a very low rate of exhumation in final diapir + erosional uplift.



*Keywords:* UHP metamorphism; Melting; Exhumation; Model; Kokchetav Massif

## 1. Introduction

At present, there is no doubt that crustal rocks subducted to greater than 100 km depths have returned to the earth's surface. However, the mechanism of exhumation of these rocks is still enigmatic. Models that have been proposed to explain exhumation of the high-pressure (HP) rocks are comprehensively reviewed by Platt (1993), Chemenda et al. (1996) and by Dobretsov and Kirdyashkin (1998).

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Though ultrahigh-pressure (UHP) complexes have much in common, there are also some differences between them (Liou et al., 1994; Dobretsov et al., 1995). For this reason, a single model cannot explain the exhumation of all UHP complexes (Platt, 1993) or all stages of exhumation. As in physics where experiment is a criterion for deciding the correctness of theoretical models, the exhumation models must be tested for their suitability to explain the established geological and petrologic data.

The present paper is aimed at analyzing the available geological and petrological material on the UHP complex of the Kokchetav Massif to determine boundary conditions for an exhumation model of the crustal rocks subducted to a depth greater than 120 km.

Because structural aspects have been discussed by Dobrzinetskaya et al. (1994) and Theunissen et al. (2000a,b), emphasis will be placed here upon petrological and geochemical data, as well as upon specific structures of the western block and a comparison with theoretical considerations.

## 2. Geological structure

According to Dobretsov et al. (1995), the Kokchetav Massif is a megamelange composed of slices and blocks formed at ultrahigh (UHP) and high pressures (HP; units 1 and 2), medium pressures (MP; Barrovian-type metamorphism, unit 3) and low pressures (LP). The UHP and HP rocks are further subdivided into two domains (Dobretsov et al., 1998, 1999): western Kumdy-Kol and eastern Kulet (Fig. 1) which are characterized by contrasting P–T conditions of metamorphism and deformation types, suggesting different mechanisms of exhumation of the western and eastern blocks (Theunissen et al., 2000b). Ultrahigh pressure unit 1 consists of diamond-bearing metasedimentary rocks and gneisses with eclogites occurring in the core of a large antiform covered or substituted by unit 2, which consists of eclogite-bearing melange with micaschists predominating in the western and/or orthogneisses in the central parts of megamelange.

According to data from drill cores, the diamondiferous rocks in the western block occur in the form of slices in granite–gneisses whose thickness reaches up to hundreds of meters (Fig. 2). It is important that at the Kumdy-Kol deposit and at the Barchinsky site, the rocks steeply dip at 60°–80° southeastwards. In general, the rocks are of ENE strike. The dip of the rocks is traceable by boreholes to a depth of 600 m at the Kumdy-Kol deposit (Fig. 2) and 100 m at the Barchinsky site. These data indicate that the western block cannot be considered as a subhorizontal structure. Fig. 2 shows that pyroxenites, eclogites and carbonate rocks as well as biotite–garnet diamondiferous gneisses

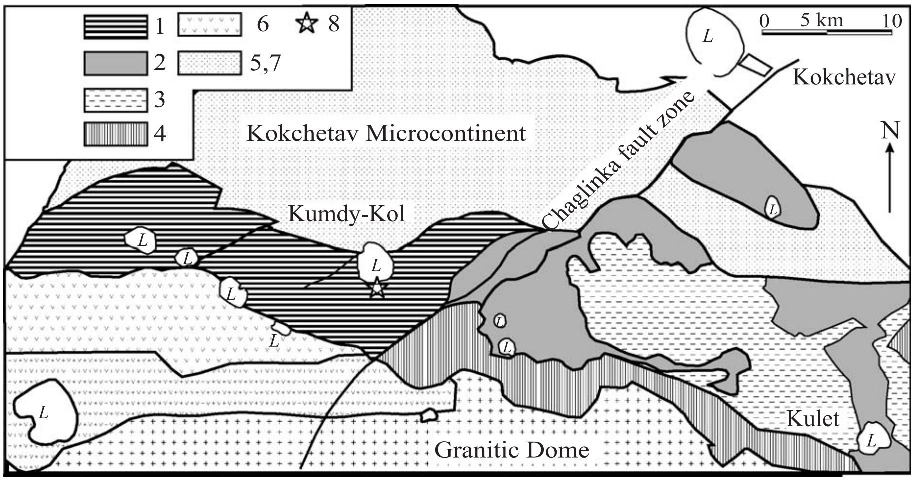


Fig. 1. Metamorphic geology map of the Kokchetav Massif with distinct western and eastern UHP domains. (1) Unit 1 UHP (diamond bearing); (2) Unit 2 UHP (coesite-bearing); (3) MP unit 3 (Al sediments and coronitic metagabbro–metadolerite); (4) LP unit; (5 + 7) microcontinental sequences (gneissic basement with its sedimentary cover); (6) Efimov unit (amphibolites, amphibole schists and quartzites); and (8) Kumdy-Kol deposit.

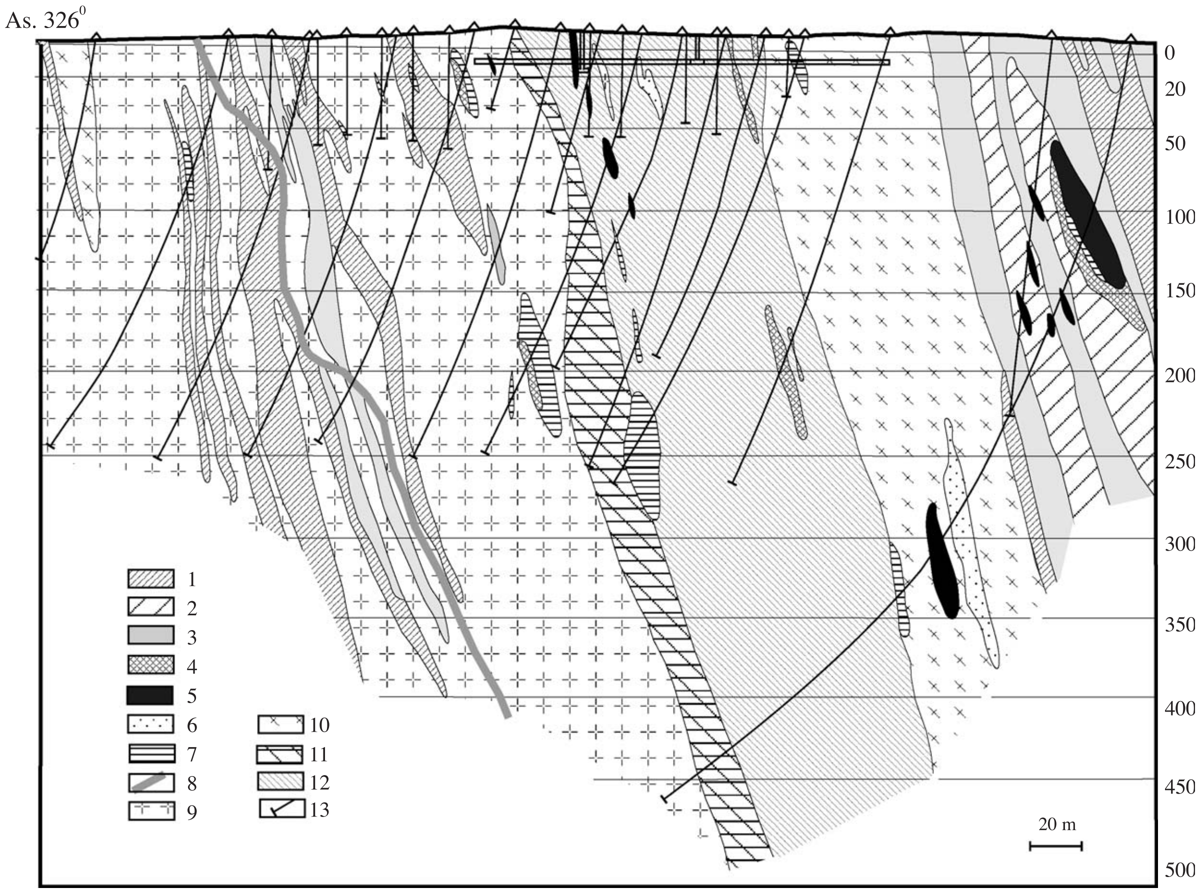


Fig. 2. Geological section of the Kumdy-Kol deposit (modified from Gonkharenco, 1990, personal communication). (1) Biotite gneisses, garnet–biotite gneisses; (2) muscovite and garnet muscovite gneisses; (3) muscovite–, garnet–muscovite and kyanite–muscovite schist; (4) dolomitic marbles; (5) eclogites, amphibolites after eclogites; (6) chlorite–actinolite–quartz rocks; (7) garnet–pyroxene rocks; (8) dykes and veins of dioritic porphiryite; (9) granite, granite–gneiss; (10) migmatites; (11) "transition zone"; (12) diamondiferous zone with biotite–, biotite–garnet graphite-bearing gneisses; and (13) drill holes.

occur in the form of lenses in granitic gneisses that contain no high-pressure phases and are most likely crystallized at midcrustal levels during exhumation. The same structure is observed at the Barchinsky site. A typical feature of the UHP rocks of the western block that distinguishes them from the UHP rocks of the eastern block is the presence of migmatite.

## 3. P–T paths

The available P–T data (Table 1) indicate that exhumation of rocks from two levels can be recognized in the Kokchetav Massif. The first corresponds to diamondiferous rocks of the western block ($P$>43 kbar, $T\sim$ 1000 °C; Fig. 3). For the eclogites and country rocks of the eastern block, peak metamorphic conditions were 650–800 °C at $P$ = 26–28 kbar, and for amphibolites, 650 °C at 10 kbar (Ota et al., 2000). There is reason to conclude that amphibolites were formed in retrograde conditions (Shatsky and Sobolev, 1993; Hermann et al., 2001).

Pyroxene–plagioclase symplectites, which developed after omphacite in eclogites of the Kumdy-Kol deposit, show that these rocks experienced granulite-facies P–T conditions (T = 800–820 °C, P = 10 kbar;

Table 1
P–T conditions of western and eastern domains

| Metamorphic stage | Temperature (°C) | Pressure (kbar) | References |
|---|---|---|---|
| *Western domain* | | | |
| Stage I | 950–1000 | >40 | Sobolev and Shatsky, 1990; Shatsky et al., 1995; Zhang et al., 1997; Hermann et al., 2001; Katayama et al., 2001; De Corte et al., 2001 |
| Stage II | >800 | 25 | Ogasawara et al., 2000 |
| Stage III | 800–820 | 10–12 | Shatsky and Sobolev, 1985; Hermann et al., 2001; Sobolev et al., 2001; Katayama et al., 2001 |
| Stage IV | 650–700 | 7–8 | Shatsky et al., 1993; Zhang et al., 1997; Hermann et al., 2001 |
| Stage V | <350 | | Shatsky et al., 1993 |
| *Eastern domain* | | | |
| Stage I | 650–800 | 27 | Shatsky et al., 1993, 1998 |
| Stage II | 600–700 | 7–10 | Ota et al., 2000 |

Shatsky and Sobolev, 1985). This is also confirmed by investigation of inclusions in zircons from diamondiferous rocks (Hermann et al., 2001). Granulite-facies conditions are also inferred from pyroxene–spinel symplectites developed around grossular–pyrope garnets in dolomite marbles and from the occurrence of sapphirine and corundum in the dolomite marbles (Sobolev et al., 2001). Newly formed pyroxene is characterized by a high content of the Ca-Tschermak component (up to 23%).

At the same time, the association Grt + Bt + Pl + Kfsp + Qtz + Cpx is common in diamondiferous biotite gneisses, in which case, the absence of orthopyroxene, which should be anticipated as a product of the reaction Bt + Pl + Qtz = Opx + Grt + Melt, must be explained. According to experimental data (Vielzeuf and Montel, 1994), at pressures above 18 kbar, the orthopyroxene-in and biotite-out curves intersect, and at high pressures, biotite can react with plagioclase and quartz without formation of orthopyroxene according to the reaction Bt + Pl + Qtz = Grt + Kfs + Melt. On exhumation, this reaction is driven to the left and we obtain the observed association of diamondiferous biotite gneisses. However, a drop in temperature from 1000 to 850 °C at about 20 kbar explains the absence of orthopyroxene during the initial stages of exhumation.

Ernst and Peacock (1996) remark that the estimated parameters of metamorphism of high-pressure complexes do not agree with the temperatures of steady-state subduction zone conditions predicted by two-dimensional thermal models. The same situation is observed in the Kokchetav Massif (Fig. 3). These authors consider a few scenarios to explain this inconsistency: heating of cold-subducted rocks after their detachment from the down-going plate and before uprise; a slow rate of convergence owing to buoyancy of continental crust entering the zone of subduction; and high-pressure metamorphism occurred prior to the attainment of steady-state conditions in the subduction zone. Nevertheless, even if all these conditions are valid, domain 1 amphibolites with an equilibrium temperature of 650–700 °C and pressure of 10 kbar (Hermann et al., 2001) do not fall within the progressive P–T trend of a subduction zone geotherm or even on the continental geotherm.

Thus, it is typical of the final stages of exhumation that granulite facies (790–820 °C, 10 kbar; Shatsky and Sobolev, 1985; Zhang et al., 1997; Hermann et al., 2001a) gives way to amphibolite-facies conditions (with biotite, phengite and amphibole) and the occurrence of migmatites.

In the discussion of exhumation models, a regressive P–T path was said to play a crucial role (Fig. 3), but plotting exhumation paths is rather difficult because in many cases, only fragments of this trend are documented. The granulite-facies stage of metamorphism is recorded only in a few specimens of eclogites and calc-silicate rocks. Based on high-Mg calcite inclusions in pyroxenes of carbonate rocks, Ogasawara et al. (2000) distinguished a stage with $T$>800 °C and $P \sim$ 25 kbar during retrograde metamorphism. Thus, during the first stage of exhumation, diamondiferous rocks were cooled from 1000 to 800 °C as pressure decreased from >42 to >25 kbar. The next stage corresponds to isothermal uplift to pressures of about 10 kbar, with subsequent nearly isobaric cooling to 600–650 °C (Hermann et al., 2001). The final stages of exhumation correspond to greenschist-facies metamorphism.

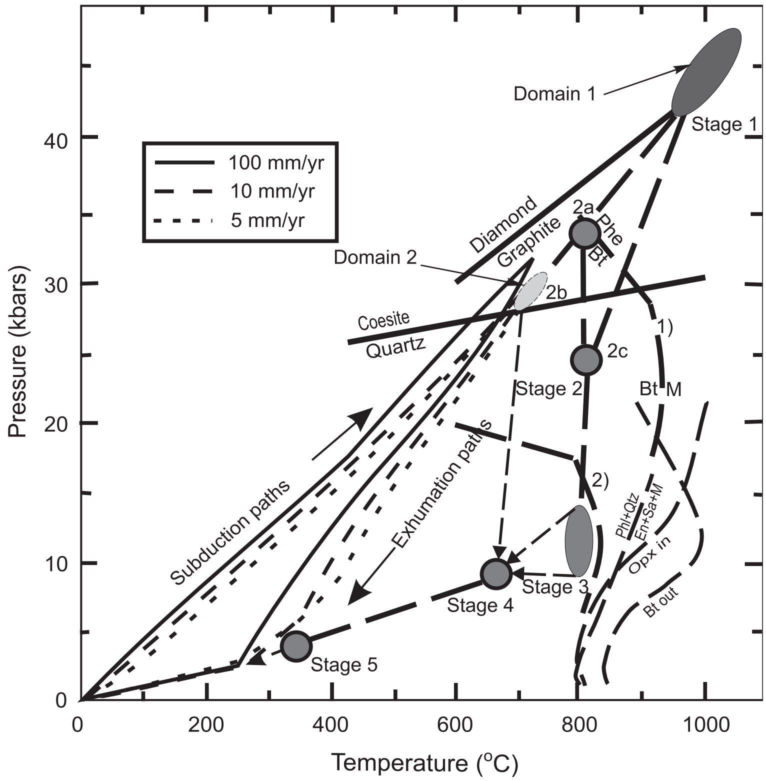


Fig. 3. P–T retrograde path of Kumdy-Kol diamondiferous rocks and Kulet domain (see text for explanation). Calculated burial and exhumation P–T paths from Ernst and Peacock (1996); Graphite–diamond transition after Bundy (1980); quartz–coesite transition after Mirwald and Massone (1980); Bt-out, Opx-in curves after Vielzeuf and Montel (1994); Phl + Qtz = En + Sa + M after Vielzeuf and Clemens (1992). (1) Bitite stability in KCMASH system (Hermann and Green, 2001); (2) biotite stability in methapelites (Vielzeuf and Holloway, 1988).

The most important question that remains to be answered is how to connect the P–T points corresponding to the peak UHP metamorphic stage ($P$>40 kbar, $T$ = 1000 °C) and retrograde granulite-facies metamorphism (10 kbar, 820 °C). In the absence of intermediate estimates, there are widely varying possibilities for plotting the P–T trend.

## 4. Evidence for melting processes

The high-equilibrium temperatures of diamondiferous rocks suggest that they may have partially melted (Shatsky et al., 1995; Hermann and Green, 2001), and geochemical data confirm this supposition (Shatsky et al., 1995, 1999). Diamondiferous rocks demonstrate a considerable scatter in Sm/Nd values (from 0.2 to 0.96; Shatsky et al., 1999), showing that all varieties of rocks, with the exception of eclogites, had similar ratios $^{143}Nd/^{144}Nd$ ($\varepsilon_{Nd}$ value of − 13,3) prior to high-pressure metamorphism. Confirmation of partial melting is given by rare-earth elements distributions: diamondiferous rocks are depleted in light REE [$(La/Yb)_N$ 0.19–1]. A trondhjemite, which cuts an eclogite body and differs from the other granitic rocks by its high Na/K ratio, has an REE pattern consistent with the equilibrium of the melt with a garnet-enriched restite namely eclogite (Shatsky et al., 1999).

As mentioned above, diamondiferous rocks are interlayered with granitic gneisses, and migmatized garnet–biotite gneisses are also observed. It could be supposed that part of the granitic gneisses and migmatite were formed by melting of diamondiferous rocks (Shatsky et al., 1999). Melt separation is possible only when a threshold degree of partial melting is reached, controlled principally by melt viscosity. In a number of subducted metapelites, there may be beds that have undergone various degrees of partial melting. In some cases, melt separation may be followed by its displacement and a mixture of restite and melt may occur.

It is also significant that age determinations indicate similar ages for diamondiferous rocks and migmatites (Shatsky et al., 1999). The age of 512 ± 5 (Borisova et al., 1995) obtained for zircons from the granitic gneisses are very similar to the Ar–Ar age of micas from diamondiferous gneisses (Shatsky et al., 1999).

## 5. Exhumation rate of the high-pressure rocks

The mineral Sm–Nd isochrons for eclogites of the Kumdy-Kol site give a middle Cambrian age (535 ± 3 Ma; Shatsky et al., 1999). An age of 530 ± 7 Ma was obtained by U–Pb dating of zircons from diamondiferous rocks (Claoue-Long et al., 1991). SHRIMP ages for individual zones of zircons containing mineral inclusions of both UHP and retrograde stages of metamorphism show that there is no systematic difference in ages between different domains in zircons that belong to different stages of metamorphism (Hermann et al., 2001). This means that the period of exhumation of UHP rocks through depths corresponding to a pressure of 6–8 kbar is within the resolution of SHRIMP dating and took no more than 6 Ma, indicating an average exhumation rate of 1.8 cm/year.

Katayama et al. (2001) report similar estimates for rate of exhumation diamondiferous gneisses of the Kumdy-Kol site on the basis of zircon studies. According to their data, high-pressure metamorphism has an age of 537 ± 9 Ma. The second stage (retrograde amphibolite facies) occurred at 507 ± 8 Ma. These values are seen to be close to Ar–Ar ages of micas (Shatsky et al., 1999). We reason that the Ar–Ar age of secondary biotite and phengite from diamondiferous gneisses (517 Ma; Shatsky et al., 1999) must reflect the greenschist rather than amphibolite-facies stage of metamorphism since the closure temperatures of the Ar-Ar system in biotite do not exceed 300–400 °C (Dobson, 1979). Under specific conditions of metamorphism and deformation, biotite can remain a closed system at very high temperature (Maurel et al., 2003). Nevertheless, chlorite inclusions occur in the rims of zircon (Katayama et al., 2001) supporting our conclusions. These data indicate that the duration of exhumation of diamondiferous rocks from the UHP peak to greenschist-facies conditions cannot have exceeded 15 Ma. However, it is crucial for exhumation models to estimate the extent to which this rate was constant.

Evidence of rapid exhumation is given by the low degree of nitrogen aggregation in diamonds of metamorphic rocks. According to De Corte (2000), these diamonds could not exist at 950 °C for more than 1 Ma and at 1000 °C for more than 0.1 Ma. Worthy of note is that, there is some uncertainty in activation energy for nitrogen in diamond to aggregate from Ib to IaA (Taylor et al., 1996). Using an activation energy 5 eV, Finnie et al. (1994) obtained residence time of 0.2 Ma for 900 °C.

Based on zoning in garnets of the Sulu–Tyube eclogites, which belong to the Kulet domain, Perchuk et al. (1998) came to the conclusion that the exhumation rate was no less than 6.2 cm/year.

The rates of exhumation at different stages can be estimated based on the above data. Stage 1 can be inferred from the degree of nitrogen aggregation in diamonds, and must be less 0.2–1 Ma. During this time interval, the rocks must have been vertically uplifted through at least 30 km. We suggest that the rate of exhumation during this superfast stage lay ranges from 20 down to 6 cm/year.

We have no information to quantify the extensional stage (see Models of exhumation). Platt and Whitehouse (1999) used U–Pb zircon absolute ages for granulites (Betic Cordillera, Spain) and Ar–Ar mica ages to calculate a 0.6-cm/year average exhumation rate in the coaxial extension stage. Based on zircon data, the complete duration of rock exhumation from metamorphism peak conditions to amphibolite grade conditions may have been less than 6 Ma.

Accordingly, the average exhumation rate up to the beginning of stage 3 was at least 1 cm/year and could have been partly much higher taking into account the time for exhumation from point 3 to point 4 (Fig. 3). The data of Perchuk et al. (1998) on garnet zoning indicate >6.2 cm/year average exhumation rate at this stage. The difference in estimates may be due to the uncertainty or to the different time episodes considered; the exhumation rate may have been 6 cm/year rate in the beginning of the second stage and 0.6 cm/year at the end of this stage.

The last and slowest stage of diapiric upwelling and erosional uplift from a depth of 33 km to the surface proceeded at an average velocity of 3.3 mm/year and continued for 10 Ma. The total time for the exhumation of the Western Kumdy-Kol domain (from stage 1 to the surface) was 14–15 Ma and to a depth of 10 km, i.e., the 300–400 °C isotherm, 9–10 Ma. These estimates are supported by isotopic data.

## 6. Models of exhumation

We now consider how the available models of exhumation fit the evidence outlined above. One of the first numerical models for superfast exhumation was proposed by Dobretsov and Kirdyashkin (1998), according to which, the viscous accretionary wedge has an output hole and provides steady long-term subduction. When a subducting plate and a continent interact in the subduction zone, Coulomb sliding should cause forces of resistance considerably exceeding the driving forces of subduction. Therefore, these authors concluded that a viscous layer must exist between the plates and that the accretionary wedge plays the role of a lubricant. A change in configuration of the wedge leads either to a decrease or to an increase in pressure, providing for steady subduction. The cases of an unsteady state of the subduction wedge were also considered, particularly, a drastic change in viscosity and dimension ratio of the base of wedge and output hole. When the output hole is plugged by a fragment of arc or microcontinent entering the subduction zone, the pressure in the wedge drastically increases which provides a high rate of backflow. An equation was derived (Dobretsov and Kirdyashkin, 1992, 1998), permitting determination of the maximum velocity of backflow at the initiation:

$$u_{\max} = Ph_0^2/2\eta l \quad (1)$$

where $u_{\max}$ is the maximum rate at the initiation; $P$, overpressure in wedge, excess relative to lithostatic pressure ($P_{\text{tot}} - P_{\text{lith}}$); $\eta$, viscosity; $l$, length of wedge; $h_0$, height of subsurface portion of wedge. For constant dimensions of the wedge ($l$ and $h_0$), the main regulators of exhumation are $P$ and $\eta$ (Fig. 4). In terms of this model, the maximum rate of return flow can reach meters per year, but the real rate is lower than 0.1–1 m/year, with a drop in pressure as a result of decreased "rigidity" of the wedge. The model of Dobretsov and Kirdyashkin (1998) is essentially similar to that of Cloos (1993) but was refined by numerical modelling.

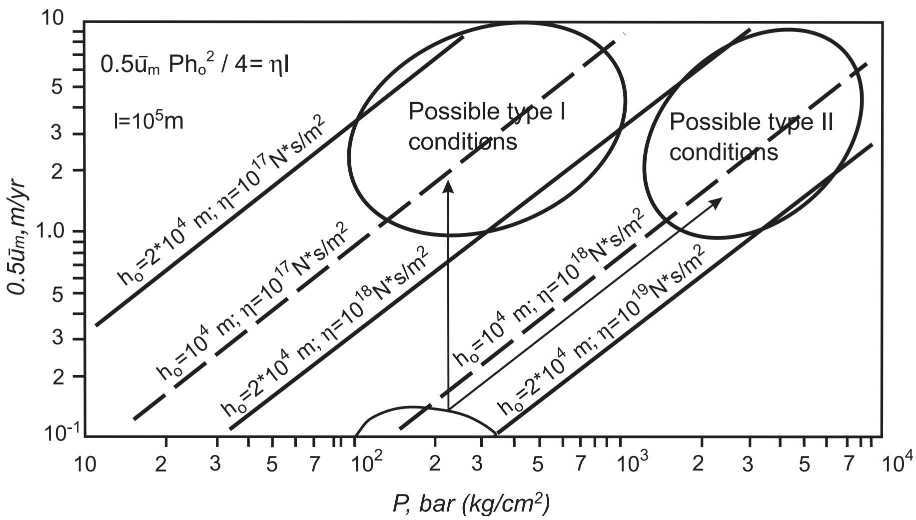


Fig. 4. Average flow velocity in an accretionary wedge versus average value of external pressure (P), viscosity ($\eta$) and wedge height ($h_0$) (Dobretsov and Kirdyashkin, 1998).

The resistance of the continental lithosphere at a depth exceeding 80–85 km drops to values of less than 100 bar (Artyushkov, 1993). Thus, this model can explain uplift of high-pressure rocks from depths not greater than 60 km and can be modified by accepting that the wedge consists of subducted crustal rocks and the viscosity inside the wedge differs drastically from the viscosity of the overlying lithosphere mantle, as discussed below. In this case, a decrease in viscosity resulting from partial melting and excess buoyancy is of crucial importance.

Chemenda et al. (1996) performed physical modelling of a three-layer lithosphere, including a plastic continental crust with strong strain weakening, ductile, a very weak lower crust and a plastic mantle layer. They established two possible regimes of continental subduction characterized by high and low pressures between the overriding and subducting plates, respectively. The subduction of continental crust reaches maximum depths for a low interplate pressure and high strength of the crust. In this case, the crust fails at the base of the overriding plate. Owing to a difference in density between crustal rocks and surrounding mantle, the subducted crust intrudes along the zone of subduction. A difference in the density between mantle and down-going continental crust as a driving force of exhumation is also considered in models by Ernst and Peacock (1996) and Ernst et al. (1997). However, these models cannot explain variation in the rate of exhumation with the depth and the P–T retrograde path of Kumdy-Kol diamondiferous rocks.

We propose the following multistage model for exhumation to account for the data available. The first stage of exhumation is initiated when a continental block (microcontinent) arrives at the subduction zone and a fragment of the crustal block is sliced off the base of the overlapping plate and then is detached from the subducting lithosphere plate, following the model by Chemenda et al. (1996). Incorporation of the thickened continental block causes instability of the subduction wedge and a backflow in it (Dobretsov and Kirdyashkin, 1992, 1998). In this case, the exhumation velocity will be governed not only by buoyancy but also by excess pressure within the wedge. As mentioned above, this cannot be great at depths exceeding 60 km. It follows from Eq. (1), however, that a decrease in excess pressure in the wedge ($P_{tot} - P_{lith}$) is compensated by a drastic decrease in viscosity resulting from partial melting.

Rapid exhumation of diamondiferous rocks along the zone of continuing subduction will favor their cooling to 800 °C (stage 2a in Fig. 3). Given P–T parameters of the eastern block, the temperature in the subduction zone at depths of about 90 km (27–30 kbar) does not exceed 800 °C. Below 800–850 °C, the proportion of melt decreases drastically which leads to increased viscosity and incorporation of a continental block that is too thick to subduct. Thus, subduction terminates, leading to doubling of the thickness of the crust. Therefore, the next stage of exhumation is characterized by adiabatic decompression (Fig. 3), corresponding to the model for extension (Platt, 1993). If further exhumation along the zone of subduction is assumed, it is difficult to understand the cause of nearly isothermal uplift.

The second stage of exhumation has been explained in many HP and UHP complexes as resulting from the collapse of the thickened crust or to an extensional tectonic stage leading to numerous thrusts and upward movement of granitic gneiss domes. The extensional tectonic stage for the rocks of the Kumdy-Kol block continued to depths of 30 km (granulite facies) but is poorly preserved because the rocks continued to ascend to form a diapir according to the model for exhumation by buoyancy forces (Platt, 1993). Diamondiferous rocks could ascend in diapirically before melt crystallized at midcrustal levels when isobaric cooling occurred to temperatures of about 600–650 °C (stage 4 in Fig. 3). Further slow uplift for several tectonic phases was controlled by erosion and sediment accumulation (e.g., Devonian molasses).

The Kulet block did not experience the superfast exhumation. The uplift from stage 2b to stage 4 (Fig. 3) corresponds to the stage of extensional tectonics that accords well with extensional structural features. The main structural feature of the Kulet block compared to the Kumdy-Kol domain is the absence of melt in Kulet UHP metamorphism.

## 7. Discussion

In our opinion, there is sufficient evidence to recognize two domains within the belt of high-pressure rocks of the Kokchetav Massif. Different P–T conditions for the peak stage as well as different styles of deformation also indicate different mechanisms of exhumation, at least for the initial stage (Theunissen et al., 2000b). In addition to the P–T exhumation path, any model for exhumation must account for the occurrence of diamondiferous rocks among granitic gneisses in the form of steeply dipping slices (60°–80°) whose thickness does not exceed a few hundred metres. This is not compatible with the model of Kaneko et al. (2000), which is based on the idea of a sandwich-like subhorizontal structure of the high-pressure complex of the Kokchetav Massif, with unit 3 at the top and unit 1 at the bottom, underlain by the low-pressure Daulet unit. Based on these data, a model for extrusion has been proposed (Maruyama and Parkinson, 2000).

The greatest difficulties arise in estimation of pressure. Earlier, we stressed that we can estimate only the lower limit of pressure using the graphite–diamond equilibrium boundary (Sobolev and Shatsky, 1990; Shatsky et al., 1995). To substantiate higher pressures (70 kbar), Kaneko et al. (2000) used experimental results by Luth (1997) on K contents of clinopyroxene. However, the partitioning of $K_2O$ between clinopyroxene and melt depends not only on pressure but also on melt composition (Harlow, 1999). Therefore, at present, there is no direct possibility to estimate pressures from K content in pyroxene.

At the same time, inclusions of magnesite found together with calcite and dolomite in zircons in carbonate rock (Shatsky et al., 1995) may be interpreted as evidence of pressures of more than 70 kbar (Sato and Katsura, 2001). The relationships of these phases remain, however, poorly understood. If zircons grow during both prograde and retrograde metamorphism (Hermann et al., 2001; Katayama et al., 2001), another reaction such as En + Dol = Di + Mgs may be responsible for the appearance of magnesite.

In our opinion, none of the above models for exhumation takes all facts presently available into account. First, UHP metamorphic complexes consist not only of silicic rocks. A large portion of their volume is made up of eclogite which should considerably increase the density of exhumed slices. In addition, these models do not take into account the rheology of high-pressure rocks. According to Stöckhert and Renner (1998), the differential stress experienced by rocks with a difference in density of 0.2 g/cm$^3$, submerged to a depth of 100 km, is about 200 MPa. The strength of the continental crust is not sufficient to bear this magnitude of differential stress at 800 °C. In this connection, the rocks should be extruded in the form of plastic mass rather than returning to the surface in the form of a coherent block, as proposed in models by Chemenda et al. (1996), Kaneko et al. (2000) and Ernst et al. (1997).

As mentioned above, the UHP rocks in the western block occur in the form of thin slices (0.1–1 km) among granitic gneisses (Fig. 2). Structural studies indicate that shear stresses for the diamondiferous rocks are exclusively low (Dobretsov et al., 1998, 1999). It was earlier shown that the diamonds included into garnets often form intergrowths with idiomorphic crystals of mica having no traces of deformation (Sobolev and Shatsky, 1990). On the other hand, the rocks from the eastern block have a well-defined lineation and microtextures indicating intense deformation (Theunissen et al., 2000b).

The mechanical behavior of high-pressure rocks of varying composition is controlled by different minerals: omphacite in eclogites, coesite in metapelites and dolomite in carbonate rocks. According to preliminary data by Stöckhert and Renner (1998), jadeite is stronger than quartz at low temperatures, but at 600 °C, they are equally strong and at higher temperatures, the strength pattern may be reversed. Data on the effect of coesite on deformation are also tentative. Thus, the upper limit of differential stress at which flow begins lies at 800 °C and on more than 5 MPa, whereas according to estimates of the lower limit, coesite behaves plastically at higher than 650 °C at a strain rate of $10^{-14}s^{-1}$. At 800 °C, aragonite begins to flow at a differential stress of lower than 1 MPa (10 bar). With all other things being equal, an increase in strain rate will increase the differential stress which is inferred from

$$\varepsilon = A\sigma^n \exp(Q/RT), \tag{2}$$

where $\varepsilon$ is strain rate; $A$, pre-exponential factor; $\sigma$, differential stress; $n$, stress exponent; $T$, temperature; $P$, pressure; and $Q$, apparent activation energy. Thus, if we do not know the strain rate, fluid content and other factors, we have great difficulty in estimating the differential stress endured by diamondiferous metamorphic rocks.

In our opinion, no strains indicated for the early stages of exhumation of diamondiferous rocks can be explained by partial melting of metapelites and granitoid rocks in the western domain of the Kokchetav Massif. At the Kumdy-Kol deposit, diamondiferous rocks are interbedded with granitic gneisses, and biotite gneisses are migmatized (Fig. 2). Diamondiferous rocks can be interpreted as lenses inside rocks that were partially remolten before or during exhumation. In this case, they can remain rigid inside the partially melted mass. The possibility of partial melting is inferred from high temperatures of equilibrium of diamondiferous rocks (950 °C) and from geochemical data (Shatsky et al., 1995, 1999).

According to experimental data, the amount of melt produced by biotite gneisses at 950 °C and 15 kbar exceeds 25% (Patino Douce and Beard, 1995). As the melt appears, the mechanism of deformation changes (Rushmer, 1995). Shear zones, in which plastic deformation is expressed, are replaced by zones of plastic flow in which the melt and sites subjected to brittle deformation occur. Shear structure is often observed in biotite–garnet gneisses (Shatsky et al., 1995). The presence of melt drastically decreases the rock viscosity and increases the density contrast between crustal and surrounding mantle rocks, as well as decrease frictional forces between the up-going block of sialic rocks, the subducting plate and the hanging wall side of the subduction zone. In addition, melting considerably decreases the strength of the sialic part of the subducting plate, initiating its separation and exhumation.

In conclusion, we should like to stress that the proposed scenario is based on the currently available facts which undoubtedly remain to be refined. Further improvement of the model for exhumation of UHP metamorphic complexes requires more precise definition of the retrograde P–T path, especially during the initial stages of exhumation. Because the rates of uplift are within the accuracy of the available methods of dating, it seems quite promising to determine the rate of exhumation on the basis of models for diffusional zoning in garnets (Perchuk et al., 1998) in the various rocks and structural units.

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