Tectonics of the Baikal Rift Deduced from Volcanism and Sedimentation: A Review Oriented to the Baikal and Hovsgol Lake Systems

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Abstract As known from inland sedimentary records, boreholes, and geophysical data, the initiation of the Baikal rift basins began as early as the Eocene. Dating of volcanic rocks on the rift shoulders indicates that volcanism started later, in the Early Miocene or probably in the Late Oligocene. Prominent tectonic uplift took place at about 20 mya, but information (from both sediments and volcanics) on the initial stage of the rifting is scarce and incomplete. A comprehensive record of sedimentation derived from two stacked boreholes drilled at the submerged Akademichesky ridge indicates that the deep freshwater Lake Baikal existed for at least 8.4 mya, while the exact formation of the lake in its roughly present-day shape and volume is unknown. Four important events of tectonic/environmental changes at about ~7, ~5, ~2.5, and ~0.1 mya are seen in that record. The first event probably corresponds to a stage of rift propagation from the historical center towards the wings of the rift system. Rifting in the Hovsgol area was initiated at about this time. The event of ~5 mya is a likely candidate for the boundary between slow and

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fast stages of rifting. It is reflected in a drastic change of sedimentation rate due to isolation of the Akademichesky ridge from the central and northern Lake Baikal basins. The youngest event of 0.1 mya is reflected by the $^{87}Sr/^{86}Sr$ ratio increase in Lake Baikal waters and probably related to an increasing rate of mountain growth (and hence erosion) resulting from glacial rebounding. The latter is responsible for the reorganization of the outflow pattern with the termination of the paleo-Manzurka outlet and the formation of the Angara outlet. The event of ~2.5 mya is reflected in the decrease of the $^{87}Sr/^{86}Sr$ and Na/Al ratios in Lake Baikal waters. We suggest that it is associated with a decrease of the dust load due to a reorganization of the atmospheric circulations in Mainland Asia. All these tectonic and climatic events could (and actually did) influence the biota of Lake Baikal. The Hovsgol rift basin was shaped to its recent form between 5.5 and 0.4 mya. However, freshwater Lake Hovsgol appeared only in the latest pre-Holocene time as a result of meltwater inflow and increase of atmospheric precipitations during the Bølling-Allerød warming. Prior to this, a significantly smaller, saline outflow-free precursor of Lake Hovsgol existed. It explains why two, now connected, lakes of similar water chemistry within similar climatic and tectonic conditions differ so much in their biodiversity.

1 Introduction

Lake Baikal is the deepest lake in the world and the largest freshwater reservoir. [Au1] It is unique not only for its size and volume (e.g., Galazii 1993), its drinkable water with low trace element composition (e.g., Suturin et al. 2003), but also for the enormous amount of endemic fauna and flora (e.g., Timoshkin 2004). Molecular dating and geological records show that evolution of biota in Lake Baikal took place on a scale of 10^4–10^6 years (e.g., Mashiko et al. 1997; Sherbakov [Au2] 1999; Koskinen et al. 2002; Hidding et al. 2003; Müller et al. 2006; Froufe et al. 2008), thanks to the long geological history of the lake and the complex history of environmental changes. Radioisotopic dating of sediments in the submerged Akademichesky ridge places the upper limit on the age of the (freshwater) Lake Baikal to as much as 8.4 million years ago (mya) (Horiuchi et al. 2003, 2004), whereas rift basins, which host Lake Baikal and its precursory lakes, are tracked back to the Middle Eocene (~45–50 mya; Logatchev and Florensov 1978; Mats 1993) or even the Late Cretaceous (~75 mya; Logachev 2003).

In Mainland Asia, the second largest freshwater reservoir is Lake Hovsgol (also written in English literature as Khubsugul, Chovsgul, or similar, and often referred to in Russian popular literature as a little brother of Lake Baikal due to the visual similarity of their environment). It is located within the same rift system (Fig. 1) with similarly severe climate conditions of long cold winters and short hot summers (Bogoyavlensky 1989; Galazii 1993). The lakes are connected through the Egin-Gol and Selenga Rivers (Fig. 1). However, Lake Hovsgol is barren in biota compared to Lake Baikal (see Table 1). A remarkable example is the absence of sponges in Lake Hovsgol, which cannot be attributed either to present-day geographic conditions or

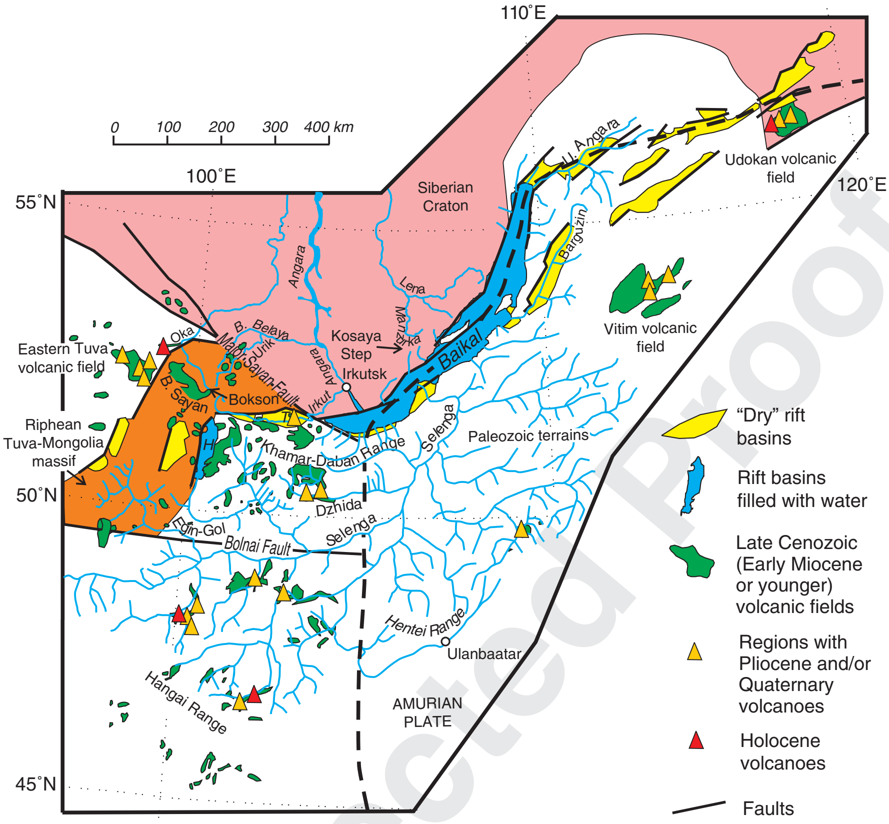


Fig. 1 Baikal rift system and surrounding regions. Rift system partially occupies a weakness zone between an ancient Siberian craton and the Pre-Cambrian Tuva-Mongolia massif with Paleozoic accreted terranes. It is limited in the south by the Bolnai Fault. The boundary of newly formed stable Amurian plate is shown by a *bold dotted line*. Its anticlockwise rotation is thought to be responsible for the opening of the Baikal basins (Zonenshain and Savostin 1981). The inflowing rover pattern to Lake Baikal is shown in detail. A few rivers mentioned in the text, not connected with Lake Baikal, namely Lena, Manzurka, Irkut, Oka, Urik, Bolshaya Belaya, are also shown. *T* Tunka basin, *H* Lake Hovsgol, *B. Belaya* Bolshaya (Big) Belaya, *U. Angara* Upper Angara, *B. Sayan* Bolshoi (Big) Sayan

water mineralization (Table 1) since sponges are known in much smaller Siberian lakes at high altitudes with higher water mineralization (e.g., Lake Chagytai; Müller et al. 2006). Some species of fauna are common to both lakes, but the fauna of Lake Baikal is more diverse (e.g., Slugina 2006).

Biota is sensitive to environmental changes, which in the past were controlled by both the climate and tectonics. The primary purpose of this chapter is to review the available information on tectonic changes in the watershed area of Lake Baikal, which includes the watershed area of Lake Hovsgol, and some adjacent regions (Fig. 1). The timing of the tectonic changes is inferred from analyses of sediments

Table 1 Comparison of Lake Baikal and Hovsgol systems

	Lake Baikal	Lake Hovsgol
Size (10^3 km^2)	31.5[a]	2.76[j]
Volume (10^3 km^3)	23[a]	0.38[j]
Average outflow (km^3/year)	57.45[b]	0.57[k]
Watershed area (10^3 km^2)	570[a]	5.13[j]
Ratio of watershed area to the lake size	18[a]	1.8[j]
Elevation (m a.s.l.)	~455[a]	~1,645[j]
Maximal depth (m)	Southern basin – 1,423[a] Central basin – 1,637[a] Northern basin – 890[a]	262[j]
Maximal thickness of sediments, m)	Southern and central basins – 7,500–8,000; Northern basin – 4,000–4,400[c]	350–450[l]
[Au3] Initiation of rift basin formations (mya)	~45–50[d], 70–75[e]?	>8[m]
Beginning of shaping the rift basins (mya)	~5 mya[f,g]	~5.5 mya[n]
Existence of fresh-water lake (mya)	>8.4[h]	~0.015[o]
Level of water mineralization (mg/l)	~150[a]	~200[j,p]
Number of animal (sub)species	>2,500[i]	<300[j,i]
Endemic animal (sub)species (%)	60[i]	~5[j,i]

Reported values have been used, or calculated from data from the following sources:
[a]Galaziy (1993)
[b]Kimstach et al. (1998)
[c]Hutchinson et al. (1992)
[d]Logatchev and Florensov (1978)
[e]Logachev (2003)
[f]Logachev and Zorin (1987)
[g]Ivanov (2004)
[h]Horiuchi et al. (2003, 2004)
[i]Timoshkin (2004)
[j]Bogoyavlensky (1989)
[k]Pisarskiy et al. (1978)
[l]Zorin (1971)
[m]Rasskazov et al. (2003)
[n]Fedotov et al. (2006)
[o]Fedotov et al. (2002, 2004)
[p]This study

within rift basins and positions of volcanic rocks on rift shoulders. In addition, some information on paleoclimate and water chemistry is also given.

2 Basic Information About the Lake Systems

Basic information about the Baikal and Hovsgol lake systems is summarized in Table 1. In general, Lake Baikal is ten times larger and has a watershed area about ten times wider. The latter is due to the significantly lower elevation of Lake Baikal compared to Lake Hovsgol. Lake Baikal is very deep and its bottom line is below sea level. Lake Hovsgol is shallower, but also deep compared to numerous other lakes in Mainland Asia.

In their classical paper, Logatchev and Florensov (1978) suggested that development of the Baikal rift basin started in the Middle Eocene with the formation of the Southern Baikal basin. Then the rift spread in two directions, western and northeastern. The initiation of the Hovsgol rift basin was placed by Logatchev and Florensov (1978) to the Pliocene–Quaternary, but more recent studies suggest that development of the Hovsgol basin started earlier, at least in the Late Miocene (Fedotov et al. 2006; see more information below on this topic). As a consequence of the early initiation of the Baikal rift basins the sedimentary infill there is much thicker compared to the Hovsgol rift basin (Table 1).

Water mineralization of Lake Baikal and Lake Hovsgol is very low; about 150 and 200 mg/l, respectively (Table 1). Elemental and strontium isotope compositions are similar except for Pb and to a lesser extent for Ba, Cu, Al, and Sr (Table 2). Baikal waters are more pristine. Probably, minor differences in water chemistry, especially in Pb, are controlled in some way by biochemical processes (see, for example, Paradina et al. 2004 for study of sponge chemistry).

3 Evidence from Sedimentary Records on Tectonic and Environmental Changes

Deepening of rift basins and growth of their shoulders (mountings), as well as climatic variations, are reflected in the rates of sedimentation and composition of the sediments associated with variations of suspended sediment and dissolved element influx by rivers and ground waters, and by variations of atmospheric precipitations. Growing mountains can also control atmospheric circulations and glaciations.

3.1 Lake Baikal

Based on analyses of sedimentary records within the Lake Baikal and Tunka basin, it was considered that rifting in the Baikal rift system had two major stages: first (pre-rift) stage of slow rifting and second stage of fast rifting (Logatchev and Florensov 1978; Logatchev and Zorin 1987). Later studies conventionally accept this subdivision into two stages for the whole rift system, but the timing for their boundary and argumentation differ. Logatchev and Zorin (1987) placed the boundary between the slow and fast rifting to ~3–4 mya. Mats (1993) described the second stage as the stage of rifting starting from the Oligocene (~25 mya), but he further subdivided this stage into several substages separated by pulses of increased tectonic activity, which correspond to about 3.5, 0.7–0.8, and 0.1 mya. A substage starting from 3.5 mya was named the Neobaikalian stage (Mats 1993). Yarmolyuk

Table 2 Minor elements, trace elements, and strontium isotope data for Lake Hovsgol, its outlet and tributary rivers sampled in anomalously dry summer of 2002 in comparison with Lake Baikal and its major tributary, the Selenga River

	MN-02–01 Hovsgol center	MN-02–06 Hovsgol north	MN-02–04 Egin-Gol	MN-02–03 Uliin-Gol	MN-02–02 Alag-Tsar-Gol	MN 02–05 Ih-Dalbain-Gol	Lake Baikal[a]	Selenga[b]
Na (mg/l)	2.9	3.1	2.4	1.2	4.8	6.7	3.3	7.1
Mg	10.9	7.3	9.1	6.1	3.7	1.2	2.6	5.5
K	0.8	1	0.8	0.4	0.6	0.8	1.02	1.4
Ca	36.1	38.1	34.1	42.1	58.1	20.0	15.7	24.3
HCO_3	151.3	151.3	146.4	161.1	207.5	90.3	N.R.	N.R.
Cl	1.1	1.1	1.8	1.1	1.1	1.1	N.R.	2.1
F	0.3	0.3	0.3	0.1	0.3	0.3	N.R.	N.R.
SO_4	2	2	4	4	4	2	N.R.	N.R.
H_4SiO_4	4	2	5	4	16	22	N.R.	N.R.
Li (μg/l)	2.9	2.8	2.8	2.0	3.9	3.8	2.1	3.4
Al	8.9	17.1	18.9	22.5	29.2	31.6	3.1	6.5
V	0.6	0.7	0.6	0.5	0.8	1.1	0.6	1.9
Mn	0.6	0.7	0.5	0.6	0.6	2.7	0.4	N.R.
Co	0.09	0.06	0.07	0.09	0.11	0.06	0.11	N.R.
Cu	2.8	6.7	2.0	2.5	3.4	2.6	0.9	1.0
Rb	0.63	0.87	0.55	0.25	0.64	0.38	0.84	N.R.
Sr	127	125	119	215	161	109	105	168
Mo	1.54	1.53	1.66	0.99	0.90	2.84	1.54	N.R.
Ba	29	29	27	39	34	22	11	14.4
Pb	2.1	6.2	1.1	1.4	1.5	1.0	0.064	N.R.
U	0.40	0.40	0.42	0.18	0.95	0.34	0.56	1.6
$^{87}Sr/^{86}Sr$	0.70869	0.70873	0.70874	0.70778	0.70953	0.70758	0.7088	0.7079

Minor elements were determined by classical "wet chemistry" at the analytical center of the Institute of the Earth's crust SB RAS. Trace elements (Li to U) were determined by inductively coupled plasma mass spectrometry using the same PlasmaQuad 2+ instrument and a similar protocol as in Suturin et al. (2003). Strontium isotopes were measured at the Institute of the Earth's crust SB RAS by thermal ionization mass spectrometry using Finnigan MAT 262 instrument. Details of analytical conditions can be found elsewhere (Ivanov et al. 2008).

N.R. – not reported

[a]Data from Suturin et al. (2003)

[b]Data from Falkner et al. (1997)

and Kuzmin (2004) used the same terminology, but placed the beginning of the Neobaikalian stage to 7 mya. This stage was subdivided into earlier and modern substages with a boundary at about 2.8 mya. Parfeevets and Sankov (2006) recognized a stage of compression at the end of the Miocene–beginning of the Pliocene (~5 mya), which could be naturally placed as a boundary between the two extension (rifting) stages of Logatchev and Zorin (1987). Ivanov (2004), using data of Horiuchi et al. (2003, 2004), noted that the major change of sedimentation on the submerged Akademichesky ridge occurred at about the same time (see later). Below, we consider borehole data for the Akademichesky ridge, because it provides the most comprehensive information on the environmental changes within Lake Baikal, and we discuss data on reorganization of the outlet-river pattern.

3.1.1 Sedimentary Record of Submerged Akademichesky Ridge

Due to international efforts, a number of deep sub-bottom boreholes were drilled as part of the Baikal Drilling Project (BDP) in winter seasons 1993–2003. The longest sedimentary record was obtained by two cores of the BDP-96-1 and BDP-98-2 boreholes drilled at the submerged Akademichesky ridge (Fig. 2). There was some

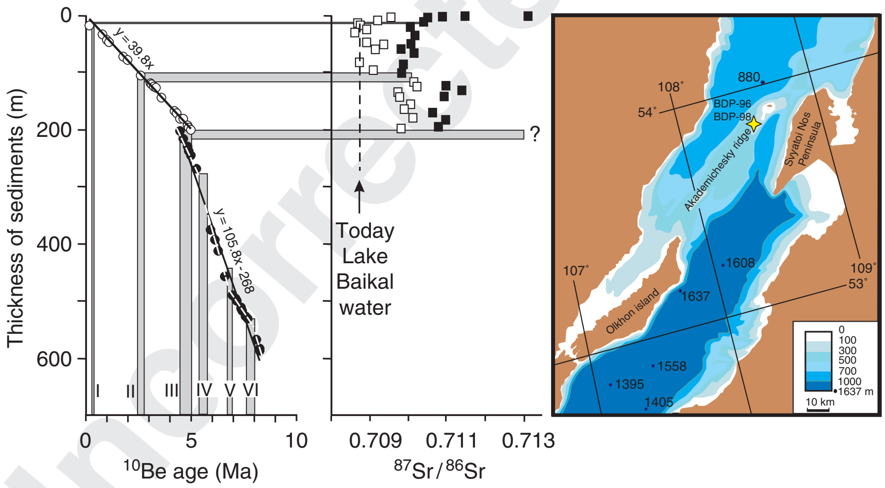


Fig. 2 Age of sediments calculated by ^{10}Be abundance in sediments versus sedimentary thickness (*left*; after Horiuchi et al. 2003, 2004), ^{87}Sr/^{86}Sr variations in terrigenic and diatom sediments (*center*; after Kuzmin et al. 2007) and bathymetry in vicinity of the submerged Akademichesky ridge (*right*). *Open* and *solid circles* are, respectively, for boreholes BDP-96-1 and BDP-98-2 taken from the submerged Akademichesky ridge. *Open* and *solid squares* are, respectively, for diatom and terrigenic sediments from the BDP-96-1 borehole. *Shadowed* areas numbered *I–VI* mark environmental changes seen from the data (see text). *Solid lines* are linear regressions for BDP-96-1 and BDP-98-2 samples. *Dotted white lines* are eye-fitted lines through subsets of the BDP-98-2 samples. *Yellow star* is the location of the BDP-96-1 and BDP-98-2 borehole sites

debate about the correct stacking of the two cores and the dating of the BDP-98-2 core, which was resolved by the ^{10}Be dating method (Horiuchi et al. 2003, 2004). The sedimentation rate of ~0.04 mm/year within the upper 200 m of the sediments (which account for about 5 mya) is very constant. However, starting from environmental change No III, deeper (older) sediments are characterized by a more variable rate of sedimentation with an average of ~0.11 mm/year. This can be explained by the rapid isolation of the Akademichesky ridge from the Central and Northern Baikal basins. Prior to 5 mya, terrigenic sediments from major rivers (e.g., Selenga, Barguzin, Upper Angara) deposited on the precursor of the Akademichesky ridge had a relatively high rate of sedimentation. After the building of the Akademichesky ridge, the influx of terrigenic sediments dropped and the rate of sedimentation also dropped (Fig. 3). Changes in the sedimentation rate also took place at about 5.5 mya (IV in Fig. 2, left), 7 mya (V in Fig. 2, left) and 8 mya (VI in Fig. 2, left). Considering these environmental changes and our hypothesis of the absence of the Akademichesky ridge prior to 5 mya, an increase of the sedimentation rate would mean an increasing rate of denudation in the watershed area, which can be related to rapid tectonic growth of mountains. The rate of sedimentation increased at environmental change No V (7 mya), which probably marks the beginning of the Neobaikalian stage in terms of Yarmolyuk and Kuzmin (2004).

BDP-96-1 core is represented by laminated layers of terrigenic and diatom-rich sediments. The former layers are characterized by a higher $^{87}Sr/^{86}Sr$ ratio compared to diatom-rich sediments, because diatoms gain strontium from the water, which is generally less radiogenic than dominant ancient (Paleozoic and Pre-Cambrian) rocks sampled by incoming rivers (Kuzmin et al. 2007). $^{87}Sr/^{86}Sr$ ratios in the most recent sediments are higher compared to present-day Lake Baikal water (Fig. 2) because terrigenic sediments are admixed with diatoms at variable proportions.

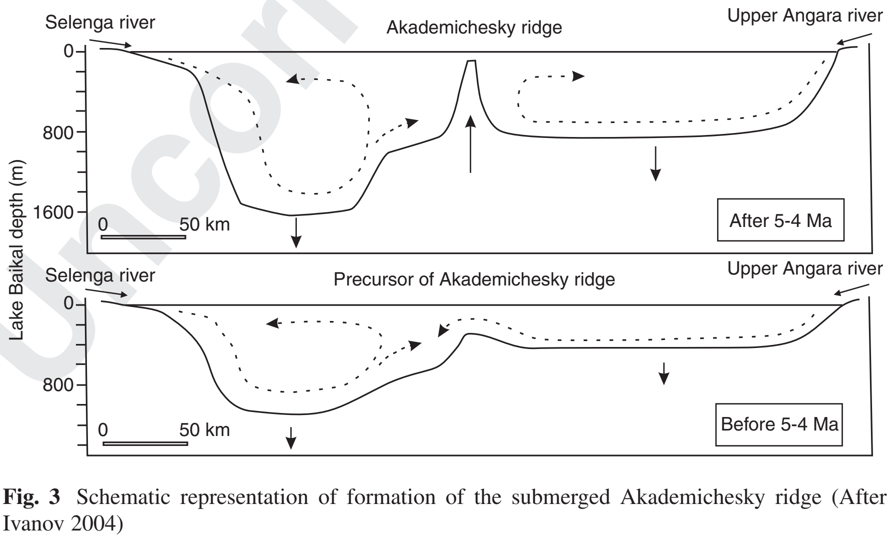


Fig. 3 Schematic representation of formation of the submerged Akademichesky ridge (After Ivanov 2004)

About 2.5–2.8 mya (II in Fig. 2, left), $^{87}Sr/^{86}Sr$ ratios in both the terrigenic and diatom-rich sediments dropped significantly and then increased again to higher values in more recent times (I in Fig. 2, left). At environmental change No II, the Al/Na ratio of sediments also changed (Al_2O_3/Na_2O increased from about 0.1–0.15 at a depth of about 100 m) (Yarmolyuk and Kuzmin 2004). Strontium isotope (and other chemical indices) variations at 2.5–2.8 mya are not reflected in the sedimentation rate. Hence, they are not due to any tectonic changes in the watershed area, but rather to climatic changes (in addition to this argumentation, a tectonic uplift within the Lake Baikal watershed area and resulting increase of erosion-related influx of terrigenic sediments would increase the $^{87}Sr/^{86}Sr$ ratio in BDP samples, while the opposite is observed). As noted by the Stratigraphy Commission of the Geological Society of London, the Gauss-Matuyama magnetic chron boundary (~2.6 mya) is the natural beginning of the Quaternary, instead of the Pliocene–Quaternary boundary at 1.8 mya (see Gibbard et al., 2005). Here, we suggest that the decrease of the $^{87}Sr/^{86}Sr$ ratio at about 2.5–2.8 mya reflects the change of atmospheric circulations in mainland Asia. A study of the Chinese Loess Plateau for the last 7 mya revealed a drop of $^{87}Sr/^{86}Sr$ ratio in acid-insoluble residue of sediments from 0.7230 to 0.7223, which occurred at about 2.5 mya, and a further decrease of this ratio to the present-day ratio of 0.7182 (Wang et al. 2007). This has been attributed to sorting of dust particles by the East Asian winter monsoon (of intracontinental origin), whose intensity increased at about 2.5 mya (Wang et al. 2007). The latter can be related with the growth of the Tibetan Plateau that occurred at about the same time, which prevented penetration of Indian monsoons into Mainland Asia (Wan et al. 2007).

The recent increase of $^{87}Sr/^{86}Sr$ ratio in both terrigenic and diatom-rich sediments of the Akademichesky ridge was assumed to be due to a catastrophic earthquake like those which formed Proval Bay in 1861 (Kuzmin et al., 2007). However, as we argue in a section below, this could also reflect a rapid growth of mountains as the result of postglacial rebounding.

3.1.2 Paleo-Outlets of Lake Baikal

The present-day outlet of Lake Baikal, which is the uppermost branch of the Angara River (Fig. 1), was formed fairly recently in the Late Pleistocene (e.g., Kononov and Mats 1986). The time of formation of the Angara source was estimated by molecular dating on gammarid populations as ca. 60 Ka (Mashiko et al. 1997), *Thymallus* (Grayling) as ca. 225–325 Ka (or 100–500 Ka if overall error is included; see Koskinen et al. 2002), and *Brachymystax lenok* as ca. 50–400 Ka (Froufe et al. 2008). Prior to this time, the outlet was by the so-called paleo-Manzurka River, which connected Lake Baikal with the Lena River (e.g., Logachev et al. 1974) (Fig. 1). Timing for this paleo-outlet system was estimated through palinological analysis of alluvial deposits of the paleo-Manzurka River combined with thermoluminescence (TL) dating (Trofimov et al. 1995). In earlier studies, the time of termination of the paleo-Manzurka River outlet was estimated as the Middle Pleistocene, some 200 Ka (e.g., Logachev et al. 1974; Mats 1993).

On the basis of palinology combined with thermoluminescence dating, Trofimov et al. (1995) have extended the paleo-Manzurka River outlet up to the Late Pleistocene (TL dates of 78 ± 20 Ka and 133 ± 30 Ka). All studies agree that initiation of the paleo-Manzurka River outlet took place in the Early Pliocene (before ca. 4 mya as estimated by correlations of palinocomplexes of the paleo-Manzurka and volcanic regions of northern Mongolia). It should be noted that the results of Trofimov et al. (1995) withdraw a suggestion about a transitory Middle Pleistocene paleo-outlet with a candidate locality between southern Baikal and the Irkut River (Kononov and Mats 1986, and later works of these authors) (Fig. 1).

The question is why the paleo-Manzurka River terminated? Romashkin and Williams (1997) (following Mats 1993 and his earlier works on Lake Baikal terraces) suggested water level fluctuations due to climatic changes with a drop in water level by 300 m and a rise by 150 m over the present-day level during the last 600 Ka. According to them, water level decrease started at ca. 400 Ka and reached its minimum at ca. 300 Ka. This was postulated as a reason for paleo-Manzurka termination (in addition, formation of the Angara outlet requires rapid tectonic subsidence in the uppermost Angara current). Colman (1998) pointed out that such estimations of water level fluctuations are unrealistic. He considered that terraces were formed mainly due to tectonic uplifting of the region while climatic changes were responsible for water fluctuations in the order of no more than a few meters. If we look at the site of the paleo-Manzurka alluvial deposit dated by TL as 133 ± 30 Ka in the vicinity of the Kosaya Step village near the present-day watershed boundary (Trofimov et al. 1995) (Fig. 1) and consider its present-day elevation of 700 m, the rate of tectonic uplifting will be about 2 mm/year. Such a rate is trivial for the modern mountain framing of Lake Baikal resulting from combined effects of active tectonics and glacial rebounding (Levi et al. 2002). Thus, we concur with Colman (1998) that 100 m scale of water fluctuations with much higher water levels than today are not reasonable. Moreover, recent estimations for climate-related water level fluctuations suggest up to about a 35 m drop (not rise!) of water level from about 100 Ka to about 70 Ka (at the transition from interglacial period MIS5–MIS4) (Urabi et al. 2004). Thus, high terraces along the Lake Baikal shores are due to a tectonic uplift. Increase of $^{87}Sr/^{86}Sr$ in recent diatom and terrigenic sediments can be related with glacial rebounding, which increased the rate of mountain growth and erosion, and is at the origin of the high terraces.

3.2 Lake Hovsgol

3.2.1 Sedimentation Records

Major information on sedimentation in Lake Hovsgol came from sub-bottom borehole KDP-01 (Fig. 4) which penetrated 53 m of sediments, from a number of a few meters long gravity cores, and from seismic profiling (e.g., Fedotov et al. 2002, 2004, 2008; Kazansky et al. 2005). Paleomagnetic dating suggests that the 53 m of KDP-01 drill-core account for 1,070 Ka. If this rate of sedimentation remained

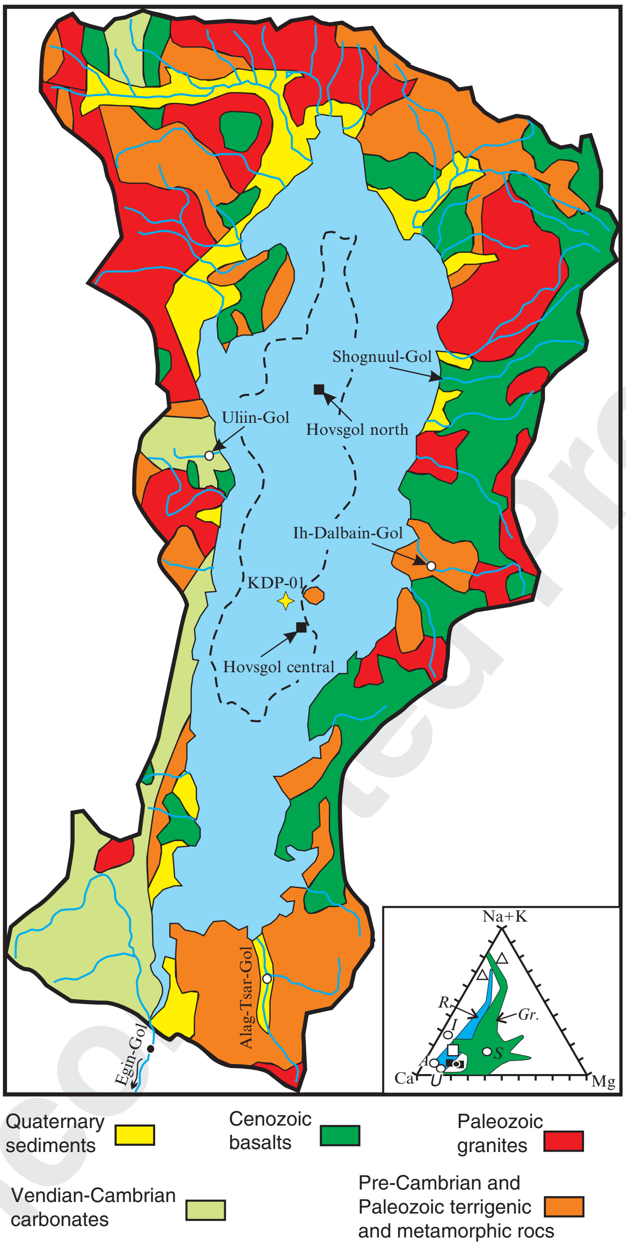


Fig. 4 Geology of the watershed area of Lake Hovsgol. *Yellow star* marks position of sub-bottom borehole KDP-01. *White circles*, *black circle* and *black squares* represent sampling sites of tributary rivers, outlet river, and Lake Hovsgol, respectively (see Table 2). *Dashed contour* in the middle of the lake marks the Late Pleistocene paleo-lake level, which was 170 m below the present-day lake level (Fedotov et al. 2004). *Insert* shows classification of river and lake waters: *white circles* are tributary rivers (*A* Alag-Tsar-Gol, *U* Uliin-Gol, *I* Ih-Dalbain-Gol, *S* Shognuul-Gol), *black circle* is an outlet river (*E* Egin-Gol), *black squares* are for Lake Hovsgol water, *white triangles* are for rain water sampled in 1969 and 1971 (Sodnom and Losev 1976), *white square* is for Lake Baikal water (Suturin et al. 2003). *Green* and *blue* areas show range of compositions measured in 1969 and 1971 for ground water (*Gr.*) and river water (*R.*), respectively (Sodnom and Losev 1976)

constant on average during the basin formation, then 350–450 m of sediments in the Hovsgol basin (Table 1) reflect that the deepest sediments are about 7–9 mya old. As will be shown in Section 4.1 this crude estimation is close to data on the beginning of rifting based on volcanism dating.

From analyses of seismic images, confirmed by gravity core data, it has been suggested that Lake Hovsgol in the latest Pleistocene was 170 m shallower compared to the present-day level (Fedotov et al. 2002, 2004) (Fig. 4). This precursory lake therefore had no outlet and was salty (Fedotov et al. 2004). The lake volume rose to the present-day volume in connection with glaciers melting and increased precipitation during the Bølling-Allerød warming (about 15–12 Ka). Analysis of the upper half of the KDP-01 core has shown that water level fluctuated several [Au4] times in the lake's history (Fedotov et al. 2008). Fedotov et al. (2004) calculated an index of tectonic activity looking at seismic images of sediments, which are derived from the angle of sediment unit deepening and number of singenetic faults in each unit. This revealed a highly active tectonic phase from 5.5 to 0.4 mya with peaks of activity at about 3 and 1 mya.

3.2.2 Strontium Balance in Water

Strontium isotope composition of water in Lakes Baikal and Hovsgol represent a mixture of strontium derived from atmospheric precipitations and terrigenic sources in the watershed area. Terrigenic sources are highly heterogeneous and depend on rock age and composition, as well as on the degree of chemical leaching of strontium from the rocks by natural waters. Climatic conditions (and hence degree of leaching) are similar for both lakes. Composition of the atmospheric precipitations is also expected to be similar, thus only composition and age of the rocks are major variable parameters. As seen from Fig. 1, rivers that flow into Lake Baikal drain mainly rocks of Paleozoic terrains, which are composed of granites, carbonate, and terrigenic rocks. Contribution from ancient granites of the Siberian craton with highly elevated $^{87}Sr/^{86}Sr$ (>0.720) and young basalts with low (~0.705) $^{87}Sr/^{86}Sr$ is small. Modeling of strontium balance for Lake Baikal was done by Falkner et al. (1997) and is not considered here. We only note that $^{87}Sr/^{86}Sr$ ratios in Lakes Baikal and Hovsgol waters are similar (Table 2).

The geology of the watershed area of Lake Hovsgol is shown in Fig. 4. Both, granites and basalts are abundant and are important sources of strontium for river waters. An additional important component is carbonate rocks of the Riphean Tuva-Mongolia massif. The insert to Fig. 4 shows that Lake Hovsgol waters are mixtures of rivers and ground waters with atmospheric precipitations. We sampled Lake Hovsgol (and the Egin-Gol River 1.5 km below its source, which is expectedly close to Lake Hovsgol in composition), Uliin-Gol, Alag-Tsar-Gol, and Ih-Dalbain-Gol Rivers in the anomalously dry summer of 2002 in a period from July 24 to August 1. We were not able to sample Shognuul-Gol, which is the only river close to ground water compositions by major compounds (insert to Fig. 4).

The contribution of different sources was modeled using the $^{87}Sr/^{86}Sr$ ratio and Sr concentrations (Faure 1986; Capo et al. 1998). First, we modeled the contribution of strontium derived from different rocks for the rivers Uliin-Gol, Alag-Tsar-Gol, and Ih-Dalbain-Gol. We discovered that Uliin-Gol waters receive 85.7%, 4%, and 10.3% of strontium from carbonates, basalts, and rain (and melted snow, which exists in high mountains in the catchment area of the Uliin-Gol through the summer), respectively. Alag-Tsar-Gol receives 41.5% and 58.5% of strontium from metamorphic rocks and rain, respectively. Ih-Dalbain-Gol receives 9%, 10%, and 81% of strontium from granites, basalts, and rain, respectively (see captions to Fig. 5

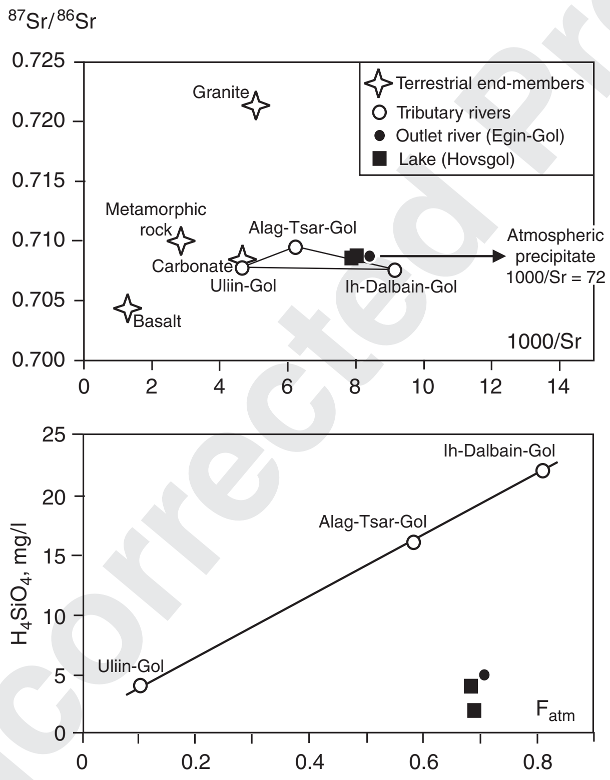


Fig. 5 $^{87}Sr/^{86}Sr$ versus 1,000/Sr (*top*) and H_4SiO_4 versus model fraction of atmospheric precipitations (*bottom*) in Lake Hovsgol, tributary, and outlet rivers. Mixing curves $^{87}Sr/^{86}Sr$ – 1,000/Sr diagram are represented by *straight lines* (Faure 1986). Average compositions for local terrestrial sources are shown: granites (Reznitskii et al. 2001; $^{87}Sr/^{86}Sr$ = 0.7214 and Sr = 200 ppm), basalts (unpublished authors' data; $^{87}Sr/^{86}Sr$ = 0.7044 and Sr = 790 ppm), carbonates (Gorokhov et al. 1995; $^{87}Sr/^{86}Sr$ = 0.7085 and Sr = 215 ppm). For metamorphic rocks, we arbitrarily set $^{87}Sr/^{86}Sr$ = 0.71 and Sr = 350 ppm. Atmospheric water composition ($^{87}Sr/^{86}Sr$ = 0.70896 and Sr = 14 µg/l) is after Sandimirov et al. (2002). The *bottom* figure shows excellent correlation between the amount of dissolved silica and the modeled fraction of atmospheric precipitations for tributary rivers. Lake and outlet-river samples do not fall on this trend due to consumption of dissolved silica in the lake by diatoms

for composition of the basalts, carbonates, granites, and metamorphic rocks, and rain waters). The ratio of the modeled contribution from different rock types is close to their ratio in the catchment area, though in the case of Uliin-Gol carbonate rocks provide even more strontium than basalts because the former are more easily leached compared to the latter. Second, we modeled the contribution of atmospheric precipitations to the total amount of strontium in Lake Hovsgol water. We found that Uliin-Gol, Alag-Tsar-Gol, and Ih-Dalbain-Gol type of waters contribute 12%, 9%, and 10% of strontium, respectively. The major source of strontium (71%) is from atmospheric precipitations. This can also be seen from the $^{87}Sr/^{86}Sr$ versus 1/Sr diagram; Lake Hovsgol waters together with Egin-Gol waters are shifted far to the right on the diagram from rivers' composition (Fig. 5, top). Calculations performed in 1969–1971 have shown that direct atmospheric precipitations to Lake Hovsgol are 48% of the total input (Sodnom and Losev 1976). This value is about 20% lower than that obtained from our modeling based on strontium data. Whether this mismatch could result from incomplete sampling or is a real feature cannot be answered at present. To decrease the amount of atmospheric precipitation in our modeling we need to assume that the contribution of strontium from rivers of the Uliin-Gol type is higher than 1/3 (as in our modeling); this, however, is unlikely. An interesting and at the moment speculative idea is that Lake Hovsgol still contains ancient ice-melted waters. Considering the average outflow rate of 0.57 km^3/year, the entire volume of Lake Hovsgol should have been completely overturned in 670 years if no evaporation was considered, and thus no ancient meltwaters from Bølling-Allerød warming could be preserved (note: Lake Baikal overturn is twice as high; Table 1). However, the present-day lake volume was not formed instantaneously at the Bølling-Allerød warming by meltwaters; it depended on variations of humidity/aridity of the climate. As pointed out by Fedotov et al. (2004), at about 5.5 Ka there was aridification of the climate, which could have decreased the Lake Hovsgol volume (thus reducing or even stopping the Egin-Gol outflow). Assuming a lower outflow rate, it seems probable that some ancient waters might exist in Lake Hovsgol. But this question requires additional studies.

Interestingly, concentrations of dissolved silica in inlet-river samples correlate with the modeled fraction of atmospheric precipitations (Fig. 5, bottom). Lake and outlet-river samples are characterized by depletion in dissolved silica; if this is not an artifact of limited sampling, it shows that silica is consumed in Lake Hovsgol by diatoms.

4 Volcanism as a Marker of Tectonic Processes

Basaltic volcanism is surface expression of melting at mantle depth (e.g., the source of melting beneath the Hovsgol region was estimated to be ~50–85 km deep; Demonterova et al. 2007). Melting is thought to be induced by upwelling from the transition zone of the mantle (410–660 km depth) of hotter and also fertile (with lower melting point) material compared to ambient mantle (Zorin et al. 2006).

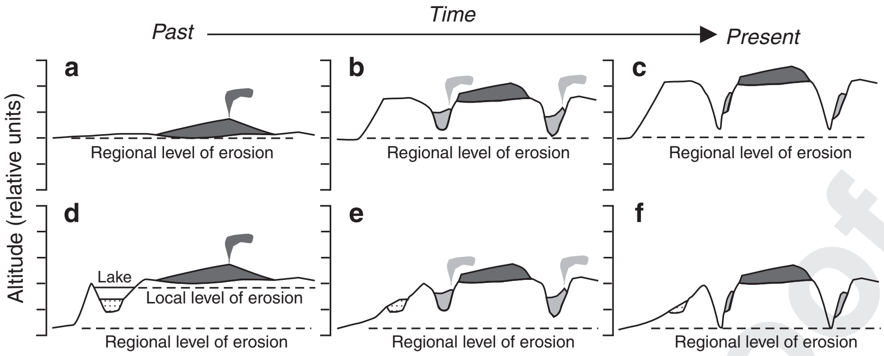


Fig. 6 A schematic model for formation of "summit" and "valley" lava remnants due to tectonic uplift at constant basal level of erosion (**a–c**) and due to sudden lowering of the regional level of erosion, in this case, as a result of a lake drainage (**d–f**). In both cases, the final effect will be the same: older lavas situated at higher elevations, usually on top of the mountains, and younger lavas occupying river terraces. *Dark* and *light gray* colors are for older and younger lavas, respectively. *Dotted area* is for lacustrine sediments

According to gravity data, several upwellings, referred to as upper mantle plumes, were the origin of high mountains; namely Hangai, Hentei, and Sayan ranges and uplands close to the Udokan and Vitim volcanic fields (Fig. 1) (Zorin et al., 2003). Thus, one may expect that rapid mountain growth should be followed by volcanic events. Schematically, this principle is shown in Fig. 6. Mountain uplift leads to fast erosion and formation of river valleys, which are filled by later lavas. It is worth mentioning that rapid river valley formation can result from the decrease of local level of erosion without any uplift, as for example in the case of paleolake drainage (Fig. 6). Such paleolakes existed in the Baikal rift system; for instance, Miocene lacustrine deposits are buried beneath lavas of the Vitim volcanic field (Rasskazov et al. 2000). Having this in mind, we focus only on those examples, where erosion could only result from tectonic uplift.

Similar argumentation was used by Rasskazov et al. (1997) in reconstruction of tectonic uplifts. Based on analysis of dated lava in relief Rasskazov et al. (1997), defined four episodes of tectonic uplifts: at about 20, 16, 8–5, and 0.7 mya. [Au5]

4.1 Dating of Volcanism

4.1.1 A Comment on Dating Methods

Volcanic rocks are dated by K-Ar and ^{40}Ar/^{39}Ar methods, which explorethe ^{40}K-^{40}Ar radioactive chain. Discussion of principles of these methods and their limitations are beyond the scope of the present review; it can be found elsewhere (e.g., McDougall and Harrison 1988; Ivanov et al. 2003; Chernyshev et al. 2006).

It should be mentioned that some laboratories (especially in earlier years; e.g., Bagdasar'yan et al. 1981) produced for Baikal rift volcanics erroneous K-Ar ages due to their laboratory procedures, and many of these erroneous ages were included in popular reviews (e.g., Whitford-Stark 1987), which are still in use in western literature. Besides this, K-Ar and $^{40}Ar/^{39}Ar$ methods can produce erroneous (beyond stated errors) ages because of some natural phenomena; K-Ar (and to some extent) $^{40}Ar/^{39}Ar$ ages for young volcanic rocks can be too old due to so-called excess argon (the closer to recent times, the more critical this problem becomes), whereas the true age of old volcanic rocks can be underestimated due to loss of radiogenic ^{40}Ar. In part, the $^{40}Ar/^{39}Ar$ method allows the controlling of these two problems and thus $^{40}Ar/^{39}Ar$ ages are generally considered as more reliable compared to K-Ar ages.

4.1.2 Brief Consideration of Timing of Volcanism in the Baikal Rift System

Published ages of volcanics erupting within and in the vicinity of the Baikal rift system were reviewed recently by Rasskazov et al. (2000). According to this review with reference to the original publication (Rasskazov 1993), the oldest (with Oligocene K-Ar ages) volcanic rocks are located in the area between the Bolshaya Belaya and Urik Rivers. We resampled and dated volcanic rocks from this area because no description of the analytical technique was provided in Rasskazov (1993). The results, which show that all volcanic rocks in this region fall into the Early Miocene, are discussed in Section 4.2.2. Lavas of probably Oligocene age have been dated in the western Hovsgol area (Ivanenko et al. 1988), but this was questioned by Rasskazov et al. (2003) (see Section 4.2.1). The Early Miocene volcanic rocks have also been dated by $^{40}Ar/^{39}Ar$ and K-Ar methods in Dzhida (22–17.5 mya), Southern Baikal (~18 mya), Hovsgol (22–16.5 mya), East Sayan (20–16.5 mya), and Eastern Tuva (~18–16 mya) regions (Rasskazov et al. 2000; 2003), showing that volcanism occurred over a vast territory of the southwestern Baikal rift at that time (Fig. 1). In different regions, different numbers of volcanic events occurred at different times. There were volcanic events in the Middle and Late Miocene period and from the Pliocene to the Quaternary. Hovsgol and East Sayan examples for the Early and Late Miocene are considered in Sections 4.2.1 and 4.2.2, respectively. Pliocene to Quaternary volcanism of Dzhida and Eastern Tuva regions are considered in Section 4.2.3.

In the northwestern Baikal rift, volcanism started later in the Middle Miocene and pulsated up to the Quaternary and Holocene in the Vitim and Udokan volcanic fields, respectively (Fig. 1). In the Udokan, volcanic episodes of ~14, 7–9, 2.5–4, ~1.7–1.8, and 0.7 mya to Holocene were recorded (Rasskazov et al. 2000; Stupak et al. 2008). In the Vitim, volcanic episodes of 13–14, 9–12, 3–4, and 0.6–1.1 mya were recorded (Rasskazov et al. 2000). Since we have no additional information on these two fields compared to that of Rasskazov et al. (2000) and Stupak et al. (2008), and because these regions are outside the watershed area of Lake Baikal (Fig. 1) we remove them from further consideration.

4.2 *Evidence from Volcanism on Tectonics*

4.2.1 Example from Hovsgol

Figure 7 represents a natural example of the concept schematically shown in Fig. 6. Late Oligocene–Early Miocene "summit" lavas in the western Hovsgol area occupy a level of 2,700 m, whereas Late Miocene "valley" lavas are located within the paleo-river valley at a lower level of 2,100 m. This means that about 600 m of uplift and erosion have happened between 21–24 and 7.8 mya. There is no strong constraint to assume that either the uplift was continuous (or in a number of rapid, but small magnitude uplifts) with an average rate of about 0.04 mm/year (~600 m/13–16 mya), which is very small (see, for example, Section 3.1.2 for comparison), or that it took place rapidly between these two events of volcanic eruptions. Paleo-river terraces with Early and Middle Miocene alluvial deposits are not observed in the area. Thus, we conclude that the 600-m uplift took place rapidly just before the volcanic event dated at 7.8 mya (Rasskazov et al. 2003). Importantly, the "valley" lava unit is displaced by the major rift-controlling fault. Two lava piles with exactly the same $^{40}Ar/^{39}Ar$ ages are displaced by the fault with an amplitude of about 400 m. There are no additional

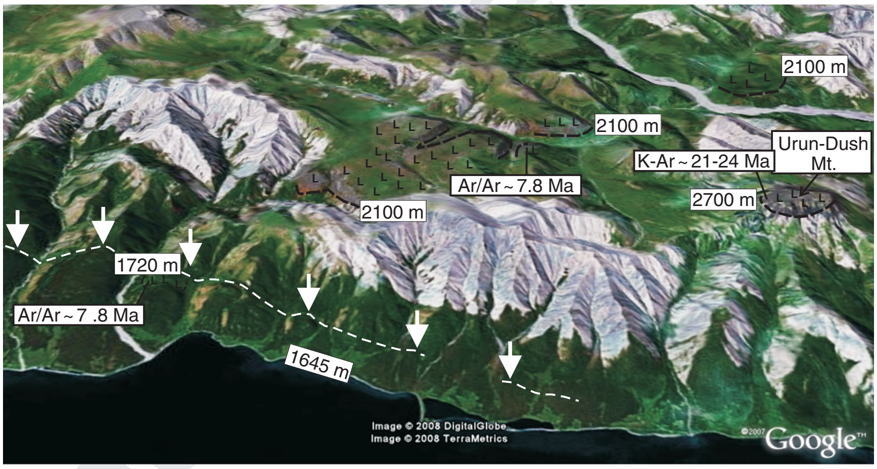


Fig. 7 A 3-D east to west view of the location of Miocene basalts (marked by "*L*" symbols) in relief of the western Hovsgol rift shoulder. *Bold dashed lines* mark the basal level of basaltic lavas, whose elevation is shown in m a. s. l.. *Tips* of *white arrows* point to a rift fault, whose surface expression is traced by the *white dashed line*. $^{40}Ar/^{39}Ar$ ages (7.76 ± 0.12 and 7.84 ± 0.06 mya for 2,100-m and 1,700-m-level lavas, respectively) have been obtained at Vrije Universiteit Brussel (Belgium) (Rasskazov et al. 2003). For a description of the analytical details see Ivanov et al. (2003). These ages are in agreement with earlier K-Ar ages for the 2,100-m-level lava unit (Ivanenko et al. 1988). However, there is some disagreement on the K-Ar dating of the 2,700-m-level lavas. Ivanenko et al. (1988) and Rasskazov et al. (2003) reported ages of 24.3 ± 0.4 (as the mean of two ages) and 21.4 ± 0.8 mya, respectively. The correct age is an important subject for a future check-up study, because this is a likely candidate for the earliest eruption in the Baikal rift system

constraints when exactly the displacement took place (this could have happened any time between 7.8 mya and the present day). However, it is logical to assume that rifting in the Hovsgol area initiated about 7–8 mya, close in time to the Late Miocene volcanic event. There is no evidence for uplift and erosion prior to 21–24 mya.

4.2.2 Examples from East Sayan

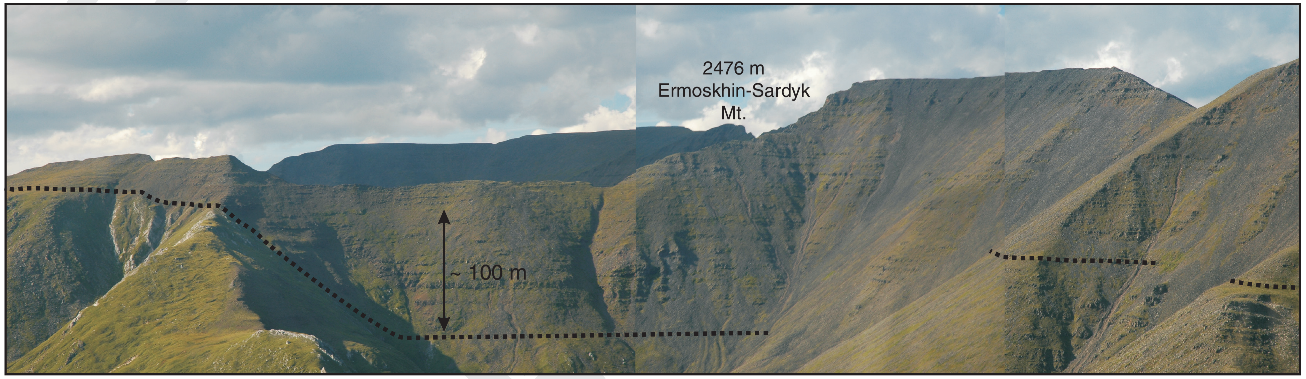
Rasskazov (1993) reported K-Ar ages of 23.7 ± 1.1 mya and 15.8 ± 0.9 mya, respectively, for the base and top of the continuous "summit" lava unit at Ermoshyn-Sardyk mounting (Urik–Bolshaya Belaya drainage area) (Fig. 1). The former age is one of the oldest among those published for the Baikal rift system and needs to be verified. It is also important to know whether these ages are correct, because, if yes, they suggest no uplift and erosion between 24 and 16 mya, which would be in [Au6] disagreement with nearby regions (Rasskazov et al. 1999). A few other K-Ar ages between 21 and 11 mya were also reported for different hypsometric levels of the Urik–Bolshaya Belaya drainage area (Rasskazov 1993).

Figure 8 (bottom) shows the position of lavas in the modern relief in this area. It may be seen that different lava units occupy levels between 2,200 and 1,500 m a.s.l. (Urik and Bolshaya Belaya Rivers are 880 and 1,030 m, respectively, at their crossing the Major Sayan Fault). However, all newly dated samples (seven in total) from five different lava units yielded K-Ar ages practically within their analytical errors at about 15–17 mya (Table 3). Neither Oligocene nor Late Miocene lavas were found, and thus we disregard the earlier K-Ar data.

Figure 8 (top) provides evidence that the Early Miocene lava filled a ~100-m-deep paleo-river valley and that the lava flows accumulated rapidly. The remnants of this valley are now at the summit elevations in the region, suggesting differentiated tectonic movements of small blocks at the boundary of two large distinct lithospheric structures; namely the Tuva-Mongolia massif and the Siberian craton. It means, first, that there was tectonic-related erosion prior to the volcanic event of 15–17 mya and, second, that much of the uplift (about 1 km) was formed after this volcanic event due to contraction at the boundary of the Tuva-Mongolia massif and the Siberian craton. Another representative example in East Sayan is a location at the Bokson River mouth, where Bokson joins the Oka River (Rasskazov 1993; Rasskazov et al. 1999, 2000) (Fig. 1). Figure 9 shows that there were two episodes of uplift and erosion; first, before the volcanic event at about 20 mya with up to 600–1,000 m of uplift and erosion, and second, before the volcanic event at about 5 mya with about 150 m of uplift and erosion.

4.2.3 Example from Khamar-Daban

Volcanism in the drainage area of the Dzhida River and at the southern end of Lake Baikal in the Khamar-Daban range started in the Early Miocene. Rasskazov et al. (2003) provided evidence that initial lava with a $^{40}Ar/^{39}Ar$ age of 21.9 ± 0.2 mya



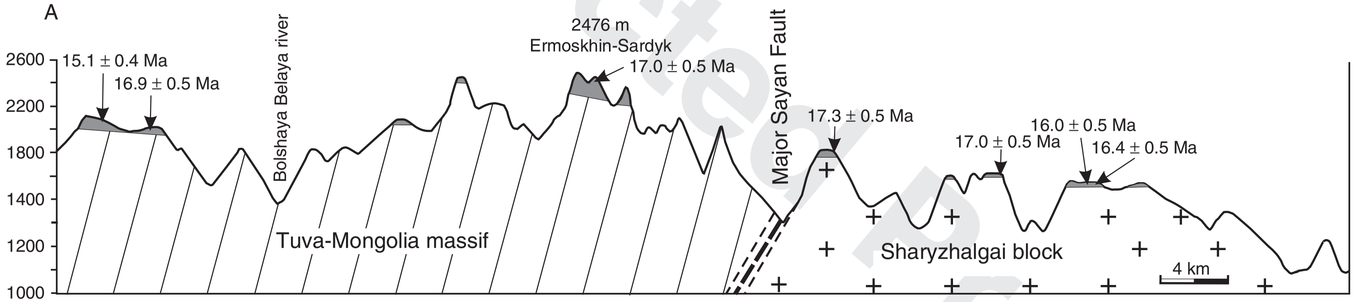


Fig. 8 A profile (*bottom*) across the Major Sayan Fault showing the position of remnants of the Early Miocene lava flows in the area between Urik and Bolshaya Belaya Rivers. Lava units are shown in *gray*. Locations of K-Ar dated samples (Table 3) are marked. A photo (*top*) gives a view on the Ermoskhin-Sardyk lava unit. *Dotted line* traces the bottom of the lava unit, showing that Early Miocene lavas buried a river valley of about 100 m deep. There is no visual discontinuity in the lava unit, providing evidence for rapid accumulation of the lava within the Early Miocene volcanic episode

Table 3 New K-Ar ages for basaltic lavas from Urik and Bolshaya Belaya watershed area, East Sayan

Sample	Location	K (%) ± σ	$^{40}Ar_{rad}$ (ng/g) ± σ	$^{40}Ar_{atm}$ (%)	Age (mya) ± 2σ
UB-07–4	52°43′52″ 101°26′05″	1.23 ± 0.02	1.407 ± 0.006	10.2	16.4 ± 0.5
UB-07–3	52°43′52″ 101°26′05″	1.44 ± 0.02	1.603 ± 0.008	10.6	16.0 ± 0.5
UB-07–9	52°42′10″ 101°23′30″	0.97 ± 0.015	1.143 ± 0.005	8.2	17.0 ± 0.5
UB-07–16	52°39′14″ 101°16′03″	1.62 ± 0.02	1.954 ± 0.008	8.8	17.3 ± 0.5
UB-07–27	52°33′18″ 100°47′31″	1.33 ± 0.02	1.568 ± 0.007	9.4	16.9 ± 0.5
UB-07–34	52°34′34″ 101°08′28″	1.55 ± 0.02	1.835 ± 0.009	11.1	17.0 ± 0.5
UB-07–40	52°34′13″ 100°45′02″	1.61 ± 0.02	1.689 ± 0.008	10.7	15.1 ± 0.4

Measurements of argon isotopes were performed at Institute of Geology of ore deposits, petrography, mineralogy, and geochemistry, Russian Academy of Sciences (Moscow) using MI-1201 IG mass-spectrometer with ^{38}Ar spike. K concentrations were determined by flame photometry at the same Institute. Analytical description can be found in Chernyshov et al. (2006). For the age calculations, conventional constants were used (Staiger and Jäger 1978).

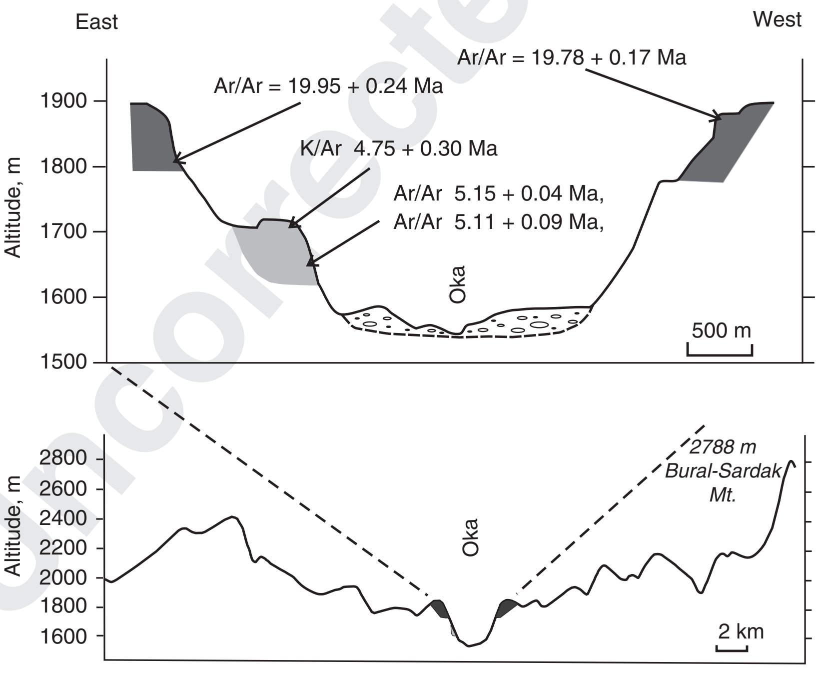


Fig. 9 A profile showing the position of the Early Miocene and Early Pliocene lava units at the site where Bokson River joins the Oka River. Lava units of different ages are shown in *grays* of different intensity. The *top* figure is reproduced after Rasskazov et al. (2000). The *bottom* figure gives an extended view to show that the Early Miocene lava erupted in an up to 600–1,000 m deep river valley

erupted on a flat surface, whereas lava with $^{40}Ar/^{39}Ar$ ages between 19 and 17.5 mya filled river valleys on both sides of the Khamar-Daban range. This places the timing of the uplift to about 20 mya.

4.2.4 Tectonically Triggered Pulses of the Pliocene and Quaternary Volcanism

The Pliocene and Quaternary volcanism occurred over the Baikal rift system and adjacent regions of Mongolia (Fig. 1); however, most of the magma volume (~87%) erupted within the Eastern Tuva volcanic field (Demonterova 2002). The reasons for this are not understood. The Pliocene–Quaternary episode of volcanism in the Eastern Tuva was preceded by two episodes of less intensive volcanism in the Early and Middle Miocene (Rasskazov et al. 2000). Here, we compare the latest pulse of volcanism in the Eastern Tuva with that in the Dzhida area, because these two regions were most thoroughly dated by K-Ar and $^{40}Ar/^{39}Ar$ methods. The Pliocene–Quaternary volcanism in the Dzhida area was also preceded by the Early Miocene volcanic event.

Published ages for the Eastern Tuva and Dzhida volcanic fields are summarized in Fig. 10 (top). A number of brief volcanic pulses are seen. Some peaks are small compared to others, reflecting different degrees of sampling of different units. With some uncertainty, we separate pulses of about 2.9, 1.9, 1.2, 0.8 (or duplicated peaks at 0.8 and 0.75 mya), and 0.6 mya for the Dzhida volcanic field and those of about 2.8, 1.75, 1.1, 0.75, 0.2 mya, and in the Latest Pleistocene–Holocene for the Eastern Tuva volcanic field (Fig. 10, top). It seems that volcanic pulses in the Eastern Tuva concordantly followed that in Dzhida with a delay of about 0.1 mya starting from the Pliocene until the Middle Pleistocene. The duration of the volcanic calms (between peaks of volcanism) in both regions shows the same correlation with the consecutive number of volcanic calms (Fig. 10, bottom). However, in the Middle Pleistocene (0.8–0.6 mya), volcanism stopped in the Dzhida area, while in the Eastern Tuva it attained another periodicity. There is no simple explanation for that. Probably, volcanic eruptions were triggered by movement of the lithospheric blocks (plates and microplates). For example, movement of the Amurian plate is thought to be at the origin of opening of the Baikal rift basins (Zonenshain and Savostin 1981) (Fig. 1). A number of other, smaller plates were also defined by GPS data (e.g., Lukhev et al. 2003). For example, ancient structures, like the Tuva-Mongolia massif, played a role in controlling extension and volcanism in the southwestern Baikal rift system (Vasil'ev et al. 1997; Demonterova et al. 2008). Movement of one plate should create additional forces on boundaries of [Au7] other plates and, in such a way, tectonic stress propagates. Here, we assume that deformations at the Amurian plate western boundary controlled periodicity of volcanism in the Dzhida volcanic field. These deformations propagated with delay to the western boundary of the Tuva-Mongolia massif and controlled volcanism in the Eastern Tuva volcanic field. This tectonic regime was a characteristic from the Pliocene until the Middle Pleistocene. At 0.8–0.6 mya, some tectonic reorganization took place, which resulted in termination of volcanism in the Dzhida area and another periodicity of volcanism in the Eastern Tuva.

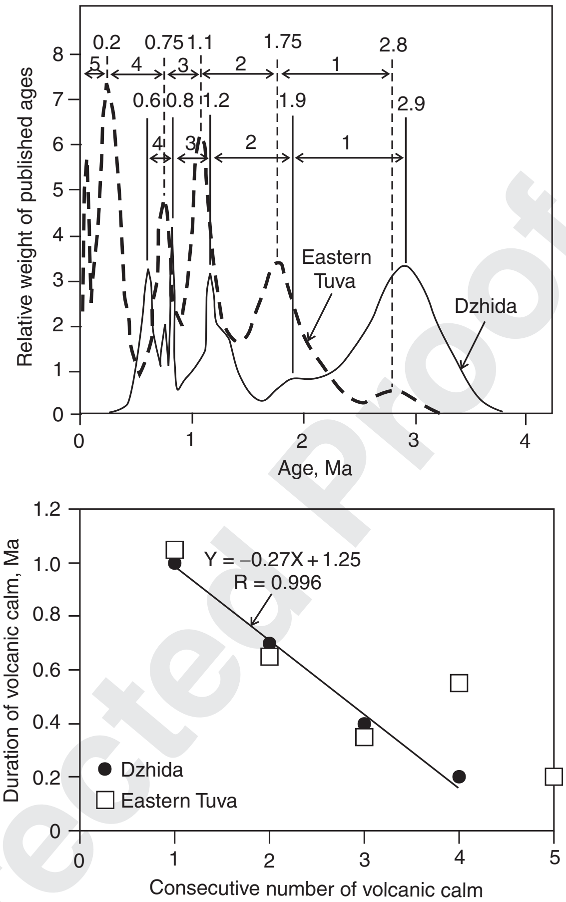


Fig. 10 Relative probability distribution of K-Ar and $^{40}Ar/^{39}Ar$ ages for the Pliocene–Quaternary volcanism in Dzhida and Eastern Tuva volcanic fields (*top*) and relation between duration of volcanic calms in respect the consecutive number of the calms (*bottom*). Age data are after Kononova et al. (1988), Rasskazov et al. (1996, 2000) and Yarmolyuk et al. (1999, 2003) (25 and 39 ages for the Dzhida and Eastern Tuva, respectively). In the *top* figure, peaks of volcanic activity are shown in millions of years (mya) and intervolcanic calms are separately numbered for each volcanic field. Height of peaks reflects a relative number of published ages and does not necessarily correspond to the amount of erupted magma

5 Discussion

The beginning of the initial stage of rifting can be placed to the Eocene or an earlier time (Logatchev and Florensov 1978; Mats 1993; Logachev 2003). However, the record of this stage is incomplete due to the rarity of outcrops with sediments of this age on the shore of Lake Baikal. The complete history of the initial stage is stored in the lowermost sediments of the Southern and Central Baikal basin, which cannot be achieved currently due to great thickness of the sediments (Table 1).

Volcanism started in the Early Miocene or probably in the Late Oligocene. The Early Miocene pulse of volcanism was widespread throughout the southwestern Baikal rift system and was preceded by tectonic uplift, which happened at about 20 mya. At this time, the East Sayan range was created and, probably, the drainage pattern in the region was completely reorganized.

The precise time when Lake Baikal came to its present-day shape is not known, but it is obvious that the lake is among the oldest on Earth. ^{10}Be dating of drill-core sediments recovered at the BDP-98-2 borehole provides evidence for at least 8.4 mya of its history (Horiuchi et al. 2003, 2004). Analysis of the rate of sedimentation suggests that there was a major change at about 5 mya when the rate decreased by about 2.5 times (Fig. 2, left). The only reasonable explanation for this is tectonic isolation of the submerged Akademichesky ridge from sedimentary input of tributary rivers. The Akademichesky ridge built up and the Central and Northern Baikal basins deepened (Fig. 3). The East Sayan range experienced an additional uplift as seen from eruption of lava into deepened river valleys at about this time (Fig. 9).

In the Late Miocene (5–8.4 mya), a few changes of the sedimentation rate are also visible in the sedimentary record of Lake Baikal. Because at this time there was no isolated Akademichesky ridge, the increasing rate of sedimentation means an increasing rate of erosion in the watershed area, which in its turn is directly related to an increasing rate of mountain growth. Such an event is seen in the sedimentary record at about 7 mya (Fig. 2, left). It is known that the Baikal rift system originated at the Southern and Central Baikal basins and propagated towards its southwestern and northeastern wings (e.g., Logachev 2003). The event of 7 mya is a likely candidate for the timing of such propagation. Data for the Hovsgol basin are in general agreement with this.

The latest tectonic event, which reorganized the outlet pattern of Lake Baikal, probably took place at about 0.1 mya. It is recorded by a rapid increase of the $^{87}Sr/^{86}Sr$ ratio in sediments which reflects an efficient proxy of terrigenic input. This event was probably, at least in part, due to glacial rebounding of the region.

Plausibly, the four tectonic events discussed here are far from complete, but these four are more pronounced and this could influence biota evolution through such mechanisms as reorganization of drainage patterns, which can control migration of animal species. For instance it has been documented that the fish species *Thymallus* (Grayling) and *Brachymystax lenok* migrated from Lena to Baikal and then to Angara and Lake Hovsgol (Koskinen et al. 2002; Froufe et al. 2008). The tectonic uplift was also at the origin of increasing rates of erosion and thus led to increased flux of terrigenous material into Lake Baikal. This could affect transparency of water and its chemical composition, and thus affect the environment. Beyond tectonics, the paleoclimate controlled biota evolution and in some cases paleoclimate could also control tectonic and volcanism patterns. For example, cessation of volcanism in the Dzhida and Vitim volcanic fields happened at about 0.6 mya and reorganization of periodicity of volcanism in Eastern Tuva took place at about 0.8 mya. Previously, we have shown that the outlet pattern of Lake Baikal reorganized at about 0.1 mya. Study of diatoms in sedimentary cores of the BDP-96-1 and BDP-96-2 has shown that *Cyclotella minuta* and *Aulacoseira baicalensis* became important in Lake Baikal starting from 760 and 120 Ka, respectively, because of interglacial episodes of global warming (Grachev et al. 1998). Is that merely a coincidence? The answer could be that climate controlled the diatom evolution, and climate controlled volcanism and tectonics at these landmarks of time.

Siberia was covered nearly completely by glaciers several times (Grosvald 1965), though the number of such glaciations and their extents are debated (e.g., Karabanov et al. 2001). Warming at the interglacial period of MIS5, known in

Siberia as the Kazantsevsky interglacial, started at about 130 Ka. It not only made the climate comfortable for diatoms such as *A. baicalensis* but also led to rapid degradation of glaciers and increasing rates of mounting growth. The latter was responsible for termination of the paleo-Manzurka outlet as argued below.

Quaternary volcanoes of the Eastern Tuva erupted under thick ice sheets several times forming specific volcanic edifices, referred to as tuya, during glacial times and forming normal lava flows and cinder cones at interglacials (Grosvald 1965; [Au8] Yarmoyuk et al. 1999; Rasskazov et al. 2000; Komatsu et al. 2004; Yarmolyuk and Kuz'min 2004). It seems that one important interglacial period in this region happened at about 0.7–0.6 mya (because K-Ar ages of 0.82–0.72 and ~0.6 mya have been obtained for tuya volcanic edifices; Yarmoyuk et al. 1999; Rasskazov et al. 2000; Demonterova 2002). Glaciers at 0.8–0.7 mya covered almost the entire Eastern Tuva volcanic field. Having in mind uncertainties of K-Ar ages for such a young period of time, the dated interglacial of 0.7–0.6 in Eastern Tuva cannot be distinguished from the interglacial derived from more accurately dated BDP cores (starting from about ~760 Ka; Grachev et al. 1998). Warming at this time created comfort conditions for *C. minuta* in Lake Baikal and stipulated the postglacial uplift in Eastern Tuva. Disappearance of the thick ice sheet led to glacial isostasy of the whole lithospheric section and probably resulted in an increase of partial melting due to decompression. It could explain why volcanism terminated in the Dzhida area, which did not experience total glaciations in the Pleistocene, and is still continuing in Eastern Tuva.

Lake Hovsgol underwent a completely different history compared to Lake Baikal. It was a small saline lake until fairly recently. In the Pre-Holocene, meltwater income and increase of atmospheric precipitations at the Bølling-Allerød warming raised water level of Lake Hovsgol to the present-day level and connected with Lake Baikal through the Egin-Gol and Selenga Rivers. This allowed new, Baikalian-like species to evolve in new ecological niches.

6 Conclusions

To summarize, we suggest that tectonic events at 20, 7, 5, and 0.1 mya could have influenced biota evolution in Lake Baikal and the surrounding regions through reorganization of drainage patterns and increasing influx of sediments into the lake. Sudden extinctions and appearances of animal and plant species as well as their degradations and flourishing recorded at other periods of time were likely associated with climatic changes and not related to tectonic activity in the Baikal rift system.

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[Au2]: Müller et al. 2006 is not in the list.
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