

Holocene tsunamis in the southwestern Bering Sea, Russian Far East, and their tectonic implications

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ABSTRACT

The Bering Sea coast of Kamchatka overlies a boundary between the proposed Okhotsk and Bering blocks, or (micro)plates, of the North American plate in the Russian Far East. A history of tsunamis along this coast for the past 4000 yr indicates that the zone north of the Kuril-Kamchatka subduction zone produces tsunamigenic earthquakes every few centuries. Such a record is consistent with convergence of the proposed Okhotsk and Bering blocks along the Bering Sea coast of Kamchatka. A tsunami deposit from the 1969 M_w 7.7 Ozernoi earthquake helps us interpret older tsunami deposits. Newly studied tephra layers from Shiveluch volcano as well as previously established marker tephra layers in northern Kamchatka provide age control for historic and prehistoric tsunami deposits. Based on >50 measured sections along 14 shoreline profiles, tsunami-deposit frequencies in the southwestern Bering Sea are about five per thousand years for tsunamis generated north of the Kuril-Kamchatka trench.

Keywords: Kamchatka Peninsula, Bering Sea, tsunamis, paleoearthquakes, tephrochronology, neotectonics.

INTRODUCTION

Plate configuration in the region of the northern terminus of the Kuril-Kamchatka subduction

zone (Fig. 1) remains an open question. Structural geology, seismicity, and global positioning system (GPS) geodesy indicate that the outermost Aleutians and the Commander Islands are moving westward with the Pacific plate along a transform boundary (Geist and Scholl, 1994; Avé Lallemand and Oldow, 2000; Gordeev et al., 2001; Bürgmann et al., 2005; Fig. 1B). Six models of the larger-scale plate-boundary configuration surrounding Kamchatka are reviewed by McElfresh et al. (2002). The simplest model assigns Kamchatka, northeastern Siberia, the Sea of Okhotsk, and the Bering Sea to the North American plate (as in Park et al., 2002; Kozhurin, 2004). However, on the basis of focal mechanisms and geologic and lineament data, Riegel et al. (1993) and Cook et al. (1986) proposed a southeastward-moving Okhotsk plate (or block) (Fig. 1A), extruded where the motion between the Eurasian and North American plates changes from extension to compression. This movement is tentatively supported by GPS data (Gordeev et al., 2001; Takahashi et al., 1999). Mackey et al. (1997) proposed that earthquake focal mechanisms indicate that an additional block, the “Bering microplate” (Fig. 1A), is slowly rotating clockwise.

If both the Okhotsk and Bering blocks are moving as proposed in these studies, then the Bering Sea coast of Kamchatka is being actively compressed, whereas if all of northeastern Russia is part of the North American plate, the southwestern Bering Sea should be tectonically quiet. Twentieth-century seismicity that supports the active-boundary model includes the tsunamigenic 1969 Ozernoi earthquake (M_w 7.7; Fig. 1C), as well as the less-documented 1945 M_w 7.3 earthquake (Gusev and Shumilina, 2004) and others in this same region (Mackey et al., 2004; Kondorskaya and Shebalin, 1982;

Fig. 1C). Another means for examining the tectonic nature of this coastline is to study its paleoseismic record via the study of tsunami deposits, which is the purpose of this paper.

Approach

Studies of tsunami-deposit records in the Russian Far East and elsewhere can make important contributions to subduction-zone and plate-motion histories. Tsunami-deposit studies on a millennial time scale fall into a plate-tectonic time frame between historical-contemporary and “geological.” The latter time frame, generally based on magnetic stripes in oceanic crust (e.g., DeMets et al., 1994), is not represented by the Holocene. The former time frame, based primarily on map compilations of historical seismicity for this region (Mackey et al., 2004; Fig. 1C) and on recent GPS measurements (e.g., Takahashi et al., 1999; Gordeev et al., 2001; Bürgmann et al., 2005), aids in understanding plate boundaries and present motions, but these records are short. Moreover, remote sites such as the Russian Far East, and submarine crust, are still hard to instrument and to track.

Along many coastlines, paleoseismological studies have aided the understanding of plate-boundary behavior, even in areas with long historical records such as Japan and Chile. Evidence of strong modern and prehistoric earthquakes and tsunamis has been found and studied along a number of subduction zones, such as Japan, North America, and Kamchatka (Nanayama et al., 2003; Kelsey et al., 2005; Pinegina and Bourgeois, 2001). In cases with patchy or short historical records, like Kamchatka and the northwestern coast of North America (Cascadia and Alaska subduction zones), paleoseismological studies provide a means for examining

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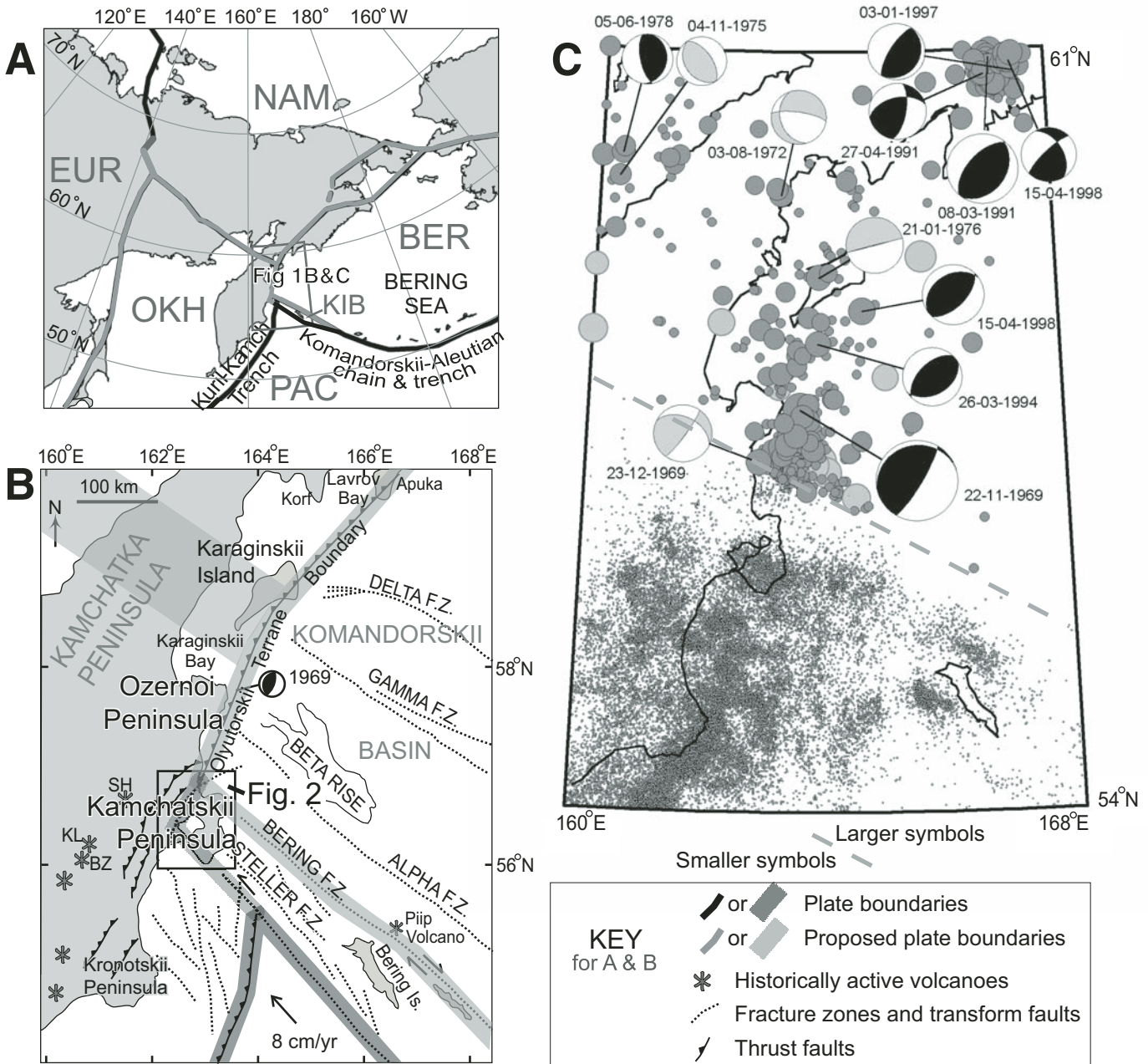


Figure 1. (A) Proposed plate and tectonic blocks in the Bering Sea region, adapted from Mackey et al. (2004). NAM—North American plate; EUR—Eurasian plate; PAC—Pacific plate; KIB—“Komandorskii [Commander] Island block” (McElfresh et al., 2002); BER—Bering block; OKH—Okhotsk block. (B) Tectonic setting of the northwest corner of the Pacific plate and western Komandorskii [Commander] Basin. Compiled and revised from Baranov et al. (1991; Komandorskii Basin); Geist and Scholl (1994); Garver et al. (2000; Olyutorskii terrane). SH—Shiveluch volcano; KL—Kliuchevskoi volcano; BZ—Bezymianni volcano. (C) Map of northeastern Kamchatka, showing seismicity and earthquake focal mechanisms. Events associated with the present-day Kamchatka subduction zone and the far western Aleutians are shown as small dots. Events north of the edge of the Pacific slab are shown by larger circles, with focal mechanisms where reliable; less well-constrained focal mechanisms are shaded gray. Focal mechanisms are shown as lower-hemisphere projections with the compressional quadrant solid. Mechanisms are taken from Savostin et al. (1983; 21-01-1976), Koz'min (1984; 04-11-1975), McMullen (1985; 03-08-1972; 23-12-1969), and Daughton (1990; 22-11-1969); all others are from the Harvard Centroid Moment Tensor Catalog. Figure 1C compiled by Kevin Mackey and Kazuyu Fujita.

recurrence intervals of large ($M_w > 7$) and great ($M_w > 8$) subduction-zone earthquakes (e.g., Pinegina et al., 2003; Atwater and Hemphill-Haley, 1997; Kelsey et al., 2005; Saltonstall and Carver, 2002; Hamilton et al., 2005).

The Pacific and Bering Sea coasts of the Kamchatka Peninsula offer a logistically challenging but outstanding opportunity to study the nature and millennial-scale history of a subduction zone and its terminus. After examining several sites along the Kamchatka subduction zone (e.g., Pinegina et al., 2003; Pinegina and Bourgeois, 2001), we treat herein the region north of that subduction zone, in the southwestern Bering Sea (Figs. 1 and 2). We will show that this area exhibits evidence of frequent tsunamis, about half as frequent as those in southern Kamchatka and more frequent than in many subduction zones, such as Cascadia. We argue that the seismicity inferred from these deposits supports a plate model with the Okhotsk and Bering blocks moving independently of the North American plate (as in Riegel et al., 1993; Mackey et al., 1997).

The site chosen for this study (the “Stolbovaia site”; Fig. 3) is apt not only for its location on Ozernoi Bay, along a disputed plate boundary, but also for its tsunami-deposit record, which we will argue is produced principally by regional earthquakes such as in 1969. Importantly, the 1969 Ozernoi tsunami deposit is well preserved and easily identified at the Stolbovaia site, because it lies directly above the 1964 Shiveluch marker tephra. By reviewing the historical tsunami catalogue, we also show that the Stolbovaia site is protected from tsunamis produced outside the Bering Sea. We use the 1969 Ozernoi earthquake and tsunami, as well as other historical tsunamis, as a guide for analyzing >4000 yr of the paleoseismic record in southern Ozernoi Bay and environs.

KAMCHATKA'S HISTORICAL EARTHQUAKES AND TSUNAMIS

To reconstruct the paleoseismicity of the Bering Sea coast of Kamchatka, we establish that tsunamis from outside the Bering Sea have not generated (most of) the tsunami deposits at Stolbovaia. Kamchatka, one of the most tectonically active regions of the world (Gorbatov et al., 1997), has a historical record of a number of large tsunamis generated along the Kuril-Kamchatka trench (Fig. 1). The 1952 M_w 9.0 earthquake was accompanied by a tsunami that completely destroyed numbers of settlements in southern Kamchatka (Zayakin and Luchinina, 1987) and that traveled across the Pacific (Soloviev and Go, 1974). Maximum observed tsunami heights onshore in the Bering Sea, however, were <2 m (Table 1). The 1737 earthquake

and tsunami (Krasheninnikov, 1755) were likely comparable to those of 1952 (Gusev and Shumilina, 2004); the reputed greater heights of the 1737 tsunami on Bering Island (Table 1) are questionable (T.K. Pinegina, 2004, unpublished field observations for Bering Island).

Kamchatka, particularly northern Kamchatka, is protected from most far-traveled tsunamis (called *teletsunamis*). For example, historical Alaskan and Aleutian earthquakes have generated minimal tsunami heights even on southern Kamchatka. The 1960 Chile tsunami is the only teletsunami with recorded large heights on Kamchatka (Table 1). The height of this tsunami in southern Kamchatka was, in general, about half the height of the Kamchatka tsunami in 1952; however, in the Bering Sea region, the two tsunamis were comparable (Table 1). Kamchatka is most susceptible to teletsunamis from South America, because tsunamis travel most efficiently in the direction of deformation (*directivity*). For older trans-Pacific tsunamis, there is no historical record for Kamchatka, but given size and directivity, the huge 1960 Chile earthquake (M_w 9.5) and tsunami are a reasonable end-member case.

Within the Bering Sea, locally generated tsunamis include postulated tsunamis produced by Aleutian volcanic landslides (e.g., Waythomas and Neal, 1998)—which would not affect our sites—and the November 1969 Ozernoi tsunami (Zayakin, 1981). The 1969 tsunami is the only historical one known to have produced tsunami heights of 5 m and more in the southwestern Bering Sea (Table 1). Historical catalogues (Soloviev and Go, 1974; Zayakin and Luchinina, 1987; Zayakin, 1996) have no record of a teletsunami or a tsunami from the Kamchatka subduction zone that generated more than ~1–2 m tsunami heights in the southwestern Bering Sea (Table 1), although the historical record is short, and coverage is spotty. The spottiness of the record increases from south Kamchatka toward the north, as in the Stolbovaia region, because of lower population densities. Hence, our tsunami-deposit studies also elucidate the historic record here.

To reconstruct the record of prehistoric tsunamis and the paleoseismicity of the southwestern Bering Sea, we need to reconstruct the general Holocene history of the coastline. Before we address these reconstructions, we establish our methods. In the following section we review the criteria for recognizing tsunami deposits, and the age control for measured sections provided by tephrochronology and radiocarbon dating.

METHODS

Stolbovaia, our field site (Figs. 2 and 3), is a narrow Holocene coastal plain stretching 25 km

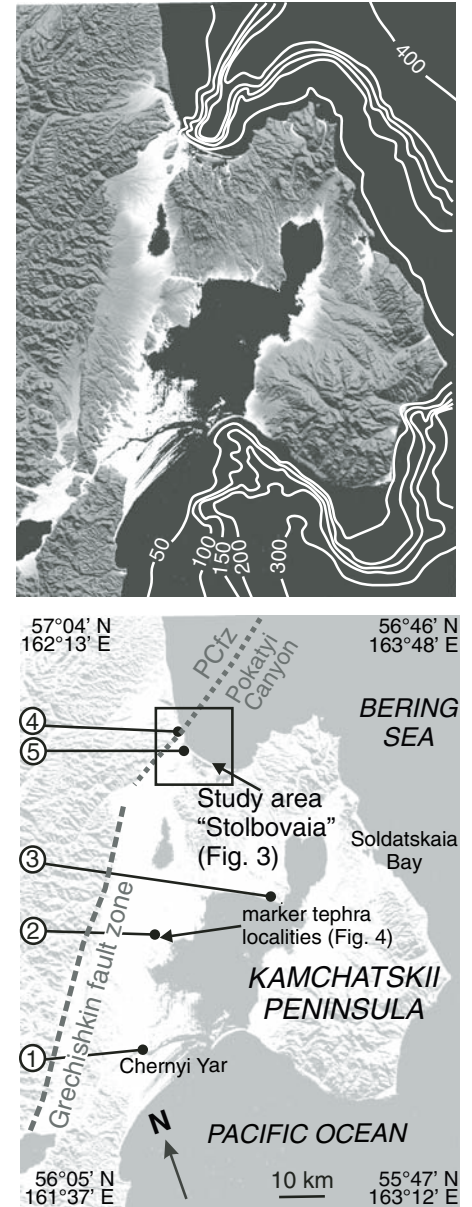


Figure 2. Above: Processed digital elevation image of Kamchatskii Peninsula (located in Fig. 1B); bathymetry (in meters) from 1:1,000,000-scale map of Kamchatka (1995, Moscow). Below: Faded image of the above, with numbered sites mentioned in the text. PCFz—Pokatyi Canyon fault zone.

along the southern coast of Ozernoi Bay, Bering Sea, between the Altyn and Stolbovaia river mouths (Fig. 3). The plain lies at the northern end of a large, north-south-oriented marshy lowland, which separates the Kamchatskii Peninsula from mainland Kamchatka (Figs. 1–3). The site is bounded at both ends by rocky headlands composed of Cenozoic volcanoclastic and marine

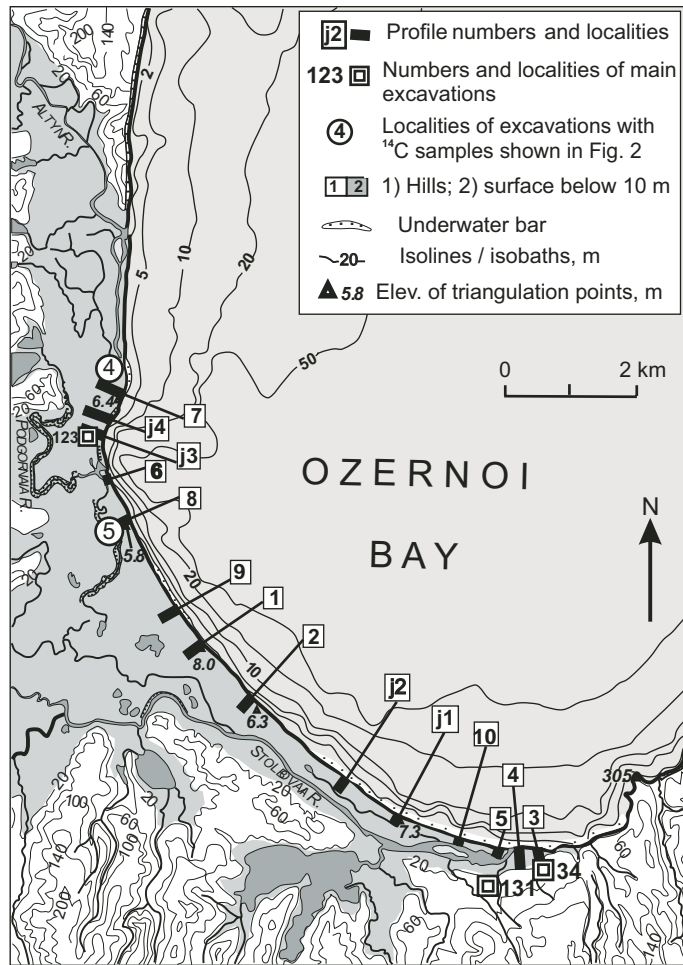


Figure 3. Topographic map of the study site (located in Fig. 2), which we call “Stolbovaia” after the Stolbovaia River, showing profiles and key excavated and measured sections.

strata and characterized by erosional landforms such as sea cliffs and marine abrasion terraces.

We measured 14 profiles (between 100 and 300 m long) perpendicular to the shoreline (Fig. 3) with a transit level and surveying rod (in a few cases with a hand level and tape). We dug >60 excavations (typically 0.5 m wide × 1 m long and 1–3 m deep), most along profiles; a few were dug in peat bogs to expose reference tephra sections, and elsewhere for reconnaissance and for exposing soils to determine ages of geomorphic features. Most (48 of 58 on-profile) excavations were at least 4 m above mean sea level (tidal range, ~1 m) and were landward of ridges 5–8 m high.

Criteria for Recognizing Tsunami Deposits

Over the past 20 yr, deposits from instrumentally and historically observed tsunamis have been studied and described (see Dawson and Shi, 2000). These studies have contributed to a general understanding of tsunami deposits and criteria for their recognition. Because each field site has specific characteristics, it is best to compare a local historical deposit with prehistoric deposits. In this study we use the 1969 Ozernoi tsunami deposit for that purpose.

Tsunami deposits are commonly sheets composed of gravel, sand, or silt eroded from adjacent beaches or other unvegetated surfaces. The layers at Stolbovaia that we interpret as tsunami deposits comprise well-rounded sand and fine gravel similar in composition and texture

TABLE 1. RECORDED OR ESTIMATED TSUNAMI HEIGHTS FROM HISTORICAL TSUNAMIS AFFECTING (OR POTENTIALLY AFFECTING) THE SW BERING SEA

Date of tsunami (local date)	Tsunami source region	Maximum tsunami height by locality (m)								Maximum anywhere on Kamchatka		
		Hilo, HI	Kamchatka just south of Bering Sea coast			Western Bering Sea coast south to north						
			Ust' Kamch tide gage	Ust' Kamch region	Bering Is.	Ozernoi Bay	Ozernoi Penin.	Korf area	Lavrov Bay		Apuka	
5 Dec 1997	Kronotskii Peninsula	0.5	n.d.								6–8	
28 Dec 1984	Kamchatskii Penin.	–	0.02								0.17	
15 Dec 1971	Kamchatskii Penin.	0.15	0.45								0.45	
23 Nov 1969	Ozernoi Peninsula	0.05	0.2	2	5–15	5–10	5–7	1–2	Effects		10–15	
4 Feb 1965	Western Aleutians	0.3									0.08	
28 Mar 1964	Alaskan Peninsula	1–3									0.06	
24 May 1960	Southern Chile	~10	0.8	3–4	3.5		2.5	>2.5	Effects		7	
5 Nov 1952	Southern Kamchatka	~3	0.1	1	2						15	
2 April 1946	Aleutians	7.9		No observations on Kamchatka; in northern Japan, 0.1–0.2 m								
14 April 1923	Off Ust' Kamchatsk	0.3		11	4						20–30	
4 Feb 1923	Kronotskii Peninsula	6.1		3							8	
28 Sept 1849	Commander Islands	–									2	
17 May 1841	Southern Kamchatka	4.6									15?	
15 April 1791	Kamchatskii Bay	–		†								
17 Oct 1737	Southern Kamchatka				30–60?						>30?	
27 Jan 1700	Cascadia	–			No observations on Kamchatka; in Japan, 1–5 m							

†Kamchatka River (Ust' Kamchatsk)—effects 7 km upstream (1791). For Kamchatka localities, see Figures 1, 2, and 3. Hilo, Hawaii, is given for calibration. Sources of data: Soloviev and Go (1974); Zayakin and Luchinina (1987); Zayakin (1996); and online catalogues.

to local beach sediments. Tsunami deposits are patchy and rarely cover most of the inundated surface. In most of our excavations, tsunami deposits thinner than 5 cm were traceable but discontinuous, partly because of bio- and cryoturbation. Tsunami deposits commonly exhibit evidence of rapid deposition, such as grading or massive (i.e., lack of) structure. All the sand layers in this study are massive to crudely graded, with no other sedimentary structures.

Tsunami deposits may share some, but not all, of their characteristics with other kinds of deposits. Storm deposits most closely resemble tsunami deposits but are not as extensive as the latter: storm wavelengths are much shorter and do not penetrate inland as far as tsunamis (e.g., Witter et al., 2003; Tuttle et al., 2004). Also, tsunami deposits generally show less contemporaneous (during the event) reworking, such as sorting and layering, than storm deposits (e.g., Goff et al., 2004).

We interpret deposits at Stolbovaia to be tsunami deposits for the following reasons. In this study we avoided excavating at elevations <4–5 m above mean sea level. All sections used for paleotsunami studies were >50 m inland from the current shoreline (at mean sea level), which, we will show, has been retreating. Thus, the deposits we interpret as tsunami deposits were left by surges of marine water that in most cases exceeded elevations of 5 m and distances of 50 m, commonly 100 m or more. We have no meteorological data for this remote locality, but in comparison to lower latitudes, the Bering Sea undergoes less intense cyclones, and the steep shelf of the southwestern Bering Sea (e.g., in comparison to the eastern Bering Sea) does not amplify storm surges. Moreover, around the Pacific, other tsunami-prone coasts with stronger cyclonic patterns, such as northern Japan, preserve tsunami deposits to the exclusion of storm deposits, at lower elevations than our excavations at Stolbovaia (e.g., Minoura et al., 1994; Nanayama et al., 2003; Kelsey et al., 2002).

Other coastal plain deposits, as well as tephra, are easy to distinguish from tsunami and storm deposits. Eolian sands are typically better (very well) sorted, finer (very fine grained sand), and form thicker, more wedge-shaped layers than tsunami deposits. Eolian deposits are present in some sections near the mouth of the Stolbovaia River. Compared to sandy tsunami and storm deposits, sand-sized tephra are better sorted, are compositionally more homogeneous (consistent with volcanic lithologies), and contain more angular grains, including crystals, cinders, and pumice. Flood deposits are typically siltier than tsunami or storm sand layers, and fluvial sediment is generally less mature than beach sediments. In our study of Stolbovaia,

few excavation localities were susceptible to river flooding or reworking.

Dating Deposits with Tephrochronology and Radiocarbon

In Stolbovaia measured sections, tephra form prominent layers that differ in thickness, color, grain size, stratification, and grading (Fig. 4; Table 2). Of 13 identifiable tephra layers, 6 are dominated by fine, light-colored ash; 5 consist of medium- to coarse-grained ash enriched in light and dark mineral grains that give them a distinctive salt-and-pepper appearance; and 2 are composed of dark-gray or black, fine to medium cinders. The characteristic appearances of tephra layers and their stratigraphic consistency help us correlate layers among excavations even without knowledge of tephra source or geochemistry.

Most marker tephra used in this study have been studied and dated elsewhere on Kamchatka and assigned averaged or rounded radiocarbon ages (Table 2; Braitseva et al., 1997a, 1997b; Pevzner et al., 1998; Ponomareva et al., 1998; Volynets et al., 1997). To use tephra layers in Stolbovaia for dating, we correlated them with previously dated sections. As a first step, we traced the tephra in a series of pits down to the Chernyi Yar peat outcrop (Figs. 2 and 4), where marker tephra have been analyzed (Pevzner et al., 1998). Then we used dispersal patterns of tephra (Fig. 5), based on data from Braitseva et al. (1997b), Ponomareva et al. (1998, 2002), and Volynets et al. (1997), to predict which marker tephra would be present at Stolbovaia and what their expected thicknesses would be. In addition, we analyzed glass shards from tephra in Stolbovaia excavations, and compared these to published data on suggested source volcanoes (Fig. 6; Table DR1).¹ On the basis of these methods, the most important tephra for the Stolbovaia site, from youngest to oldest, are SH₁₉₆₄, SH₁₈₅₄, SH₁, SH₁₄₅₀, KS₁, SH_{sp}, and SH_{dv} (Table 2; Fig. 4; Fig. 7, field example). The abbreviated coding comes from designations by Russian tephrochronologists (e.g., Braitseva et al., 1997b); e.g., SH for Shiveluch and KS for Ksudach, with subdesignations including historical dates, chronological order, or petrological characteristics.

Most tephra at Stolbovaia (Table 2) are andesitic pumiceous material from the prolific Shiveluch volcano, ~90 km WSW of the field

site (Figs. 1, 4, and 5). Fine Shiveluch ash is dominated by glass shards and has light (pale, gray, yellow, beige, tan) color, whereas coarse ash is mineral-rich and has a salt-and-pepper appearance (Braitseva et al., 1997b). Because andesitic Shiveluch tephra are similar geochemically, it is difficult to use them as markers far from the volcano on the basis either of major-element analyses of bulk samples (Braitseva et al., 1997b) or of glass-shard analyses (Fig. 6; Table DR1, see footnote 1). However, these tephra can serve as good markers locally, because in any one area (such as the Stolbovaia site) each has a characteristic grain size, color, grading, stratification, and stratigraphic position. One distinctive Shiveluch tephra, SH_{sp}, has an unusual basaltic composition (high K, high Mg) (Fig. 6; Table DR1 [see footnote 1]; Volynets et al., 1997).

In Stolbovaia, we have also identified a few other marker tephra including KS₁ (from Ksudach volcano), one of the most important Holocene marker tephra on Kamchatka (Braitseva et al., 1996, 1997b). KS₁ differs from Shiveluch tephra in composition (Fig. 6; Tables 2 and DR1 [see footnote 1]) and generally in color, and permits correlations over much of the Kamchatka Peninsula (Fig. 5). Other marker tephra at Stolbovaia come from the Kliuchevskoi volcanic group, such as gray, fine ash from the 1956 eruption of Bezymiannyi volcano (BZ₁₉₅₆; Fig. 5), which is present in northern Stolbovaia sections. The lowermost Holocene tephra recognized at Stolbovaia is a 1-cm-thick layer of black, fine- to medium-grained ash geochemically consistent with tephra from Kliuchevskoi volcano (Fig. 6; Table DR1 [see footnote 1]) and probably related to the later stage of initial cone-forming eruptions, which started ~5500–6000 yr ago (Braitseva et al., 1995).

For correlation and age control at Stolbovaia, we place more reliance on marker-tephra identification and previously determined ages rather than on our radiocarbon dates on peat, because marker tephra are well studied, but more importantly because radiocarbon dating of bulk peat generally provides ages that are younger than peat deposition. Identification of plant residue in bulk peat samples in section 4 (Podgornaia; Fig. 4; Table DR2 [see footnote 1]) shows that most peat layers, except for the lower- and uppermost ones, are dominated by the sedge *Carex cryptocarpa*, whose long roots penetrate older deposits. For this reason, the ages obtained from *C. cryptocarpa*-derived peat may be significantly younger than those measured from the remains of other plants at the same stratigraphic level. Most ages obtained for bulk peat samples from the Podgornaia sections (Fig. 4) are 100–500 yr younger than

¹GSA Data Repository item 2006047, Figures DR1–DR14 and Tables DR1–DR3, is available on the Web at <http://www.geosociety.org/pubs/ft2006.htm>. Requests may also be sent to editing@geosociety.org.

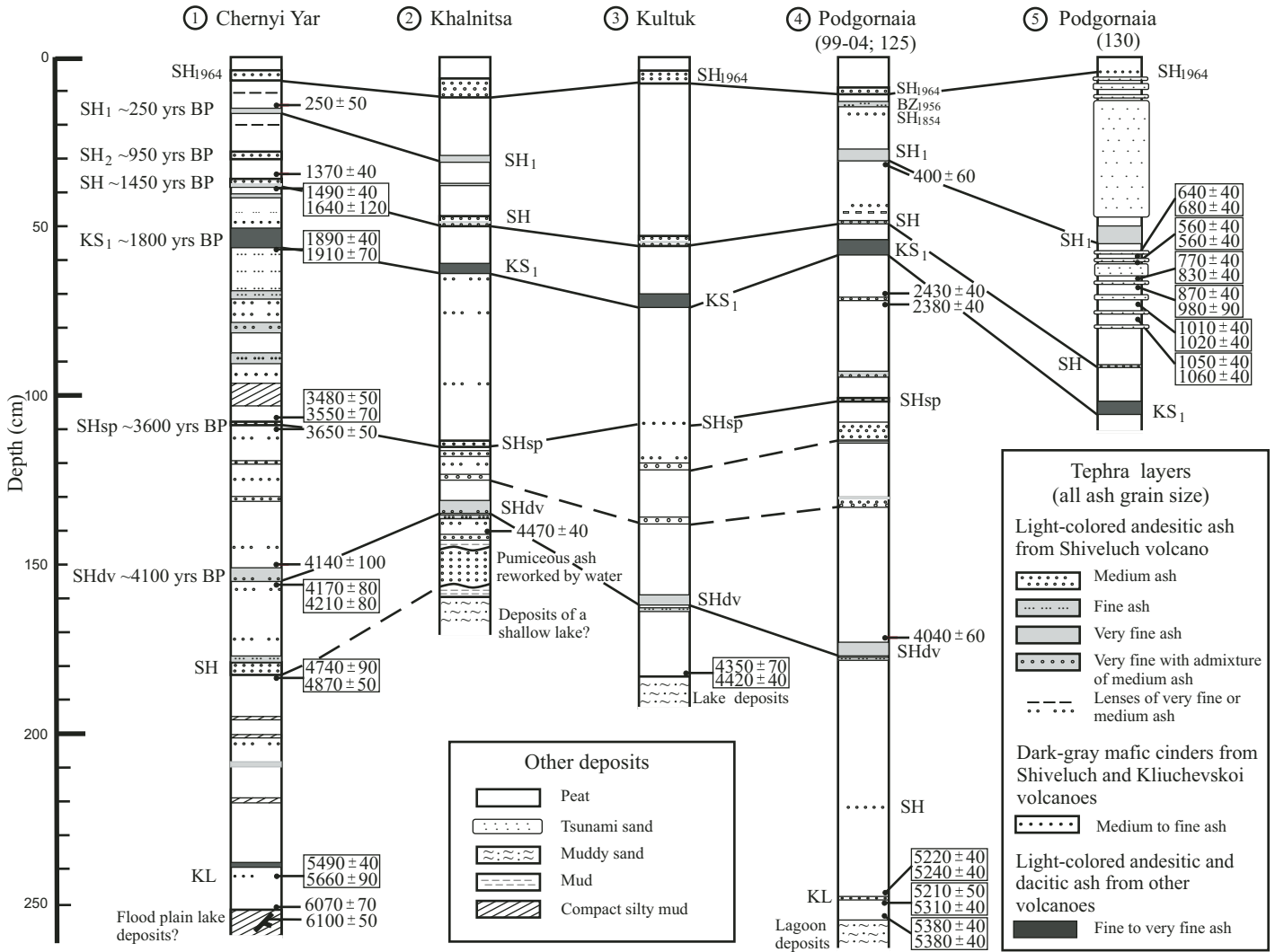


Figure 4. Stratigraphic sections of peat, organized by distance from the key reference section, Chernyi Yar, to the Stolbovaia field site (Figs. 2 and 3). Codes and sources of tephra layers and their average ¹⁴C ages are shown in Table 2. SH—Shiveluch tephras; KS—Ksudach; KL—Kliuchevskoi. Section 1 data from Pevzner et al. (1998); others are this study. Sections 4 and 5 have alternative site numbers listed, which correspond to our measured profiles. In this figure, ¹⁴C ages of marker tephras are average ages rounded to the nearest 50 yr; two dates per sample represent hot and cold extractions (see text). Lines between sections are correlations, dashed where tentative.

ages based on tephrochronology (Zaretskaia et al., 2001; Fig. DR1; Table DR2A [see footnote 1]). Reliable ages were obtained for these sections from layered peat dominated by other species of *Carex* or *Calamagrostis*; these ages are shown in Figure 4. Where the amount of material permitted, we made two successive alkaline extractions (cold and hot) from each peat sample (as in Braitseva et al., 1993).

HOLOCENE SEA-LEVEL HISTORY AND COASTAL MORPHOLOGY

To reconstruct a tsunami history for Stolbovaia, we outline the local sea-level and geomorphic history of the coastline. North

Pacific regional sea-level curves suggest a mid-Holocene sea-level high of ~2 m above the present level between ca. 7000 yr B.P. and 5000 yr B.P., followed by stabilization at or near current sea level (Douglas et al., 2001). Along the Stolbovaia coastal plain, elevations are typically 5–7 m above sea level. Below that surface, peat as thick as 3 m overlies lagoonal deposits that likely formed during the mid-Holocene highstand, which reached a local maximum at ca. 6500 yr B.P. (Melekestsev and Kurbatov, 1998). For the Stolbovaia site, we assume that for the past 4000 yr, over which we have good stratigraphic control, regional (eustatic) sea level was stable. We present evidence later for a few meters of

local sea-level fluctuation owing to subsidence and uplift.

The Stolbovaia site (Fig. 3) hosts a 30–70-m-wide beach, and in its central region (profiles 8 to J1; e.g., profile 1, Fig. 8) an active beach ridge 5–8 m high backed locally by older, more subtle beach ridges subparallel to the shoreline. Profiles 10, 5, 4, and 3 (Fig. 8) to the south, near the Stolbovaia River mouth, are variable. The area to the north (profiles J3, J4, 7) is a flat plain underlain by peat (e.g., profile 7, Fig. 9). Behind all these profiles (Figs. DR2–DR14; see footnote 1) are fluvial channels, fluviglacial deposits, and an undated and indistinct Pleistocene terrace ~10–15 m high.

TABLE 2. HOLOCENE MARKER TEPHRA LAYERS IN SOUTHERN OZERNOI BAY, NE COAST OF KAMCHATKA PENINSULA

Code	Source volcano	Ages (¹⁴ C yr B.P.)	Assigned ages (yr A.D./B.C.)	Field description	Thickness (cm)	Characteristic features
SH ₁₉₆₄	Shiveluch	Historical	1964 A.D.	White, "salt-and-pepper," medium to coarse ash	0.5–2	Medium K ₂ O content, high Cr and Sr content, presence of Hb
B ₁₉₅₆	Bezymianni	Historical	1956 A.D.	Gray, very fine to fine ash	0.5–1.5	Medium K ₂ O content, presence of Hb
SH ₁₈₅₄	Shiveluch	Historical	1854 A.D.	White, fine ash	1–1.5 (8)	Medium K ₂ O content, high Cr and Sr content, presence of Hb
SH ₁	Shiveluch	265 ± 18	1520–1690 A.D.	Pale (beige or tan), very fine to fine ash	0.5–2	Medium K ₂ O content, high Cr and Sr content, presence of Hb
SH ₁₄₅₀	Shiveluch	1450	540–640 A.D.	Yellow (beige or tan), "salt-and-pepper," fine to medium ash	1–3 (5)	Medium K ₂ O content, high Cr and Sr content, presence of Hb
KS ₁	Ksudach	1806 ± 16	160–340 A.D.	Pale yellow (beige), very fine to fine ash	3–7	Low K ₂ O content, absence of Hb
SH	Shiveluch	2400?	750–400 B.C.	"Salt-and-pepper," coarse to very coarse ash	1–1.5	Medium K ₂ O content, high Cr and Sr content, presence of Hb
SH	Shiveluch	3500	1950–1680 B.C.	Pale (beige or tan), fine to medium ash	1–1.5	Medium K ₂ O content, high Cr and Sr content, presence of Hb
SH _{sp}	Shiveluch	3600	2130–1770 B.C.	Dark-gray, fine to medium ash	1–1.5	High K ₂ O content, presence of Hb and Ph
SH	Shiveluch	3700	2280–1940 B.C.	Gray, "salt-and-pepper," medium to coarse ash	3–8	Medium K ₂ O content, high Cr and Sr content, presence of Hb
SH	Shiveluch	3800	2460–2040 B.C.	White, "salt-and-pepper," medium to coarse ash, with a fine ash top	1–1.5	Medium K ₂ O content, high Cr and Sr content, presence of Hb
SH _{dv}	Shiveluch	4105 ± 31	2870–2570 B.C.	Pale yellow (beige), fine ash	2–5	Medium K ₂ O content, high Cr and Sr content, presence of Hb
KL	Kliuchevskoi	5200	3980–4170 B.C.	Black, fine to medium ash	0.5	Medium K ₂ O content, medium Cr and Sr, absence of Hb

Note: Tephra layers are listed in chronological order. In column 3, the ages shown with error are the weighted average radiocarbon ages from Braitseva et al. (1997a, 1997b). Other radiocarbon ages are from Pevzner et al. (1998), Ponomavera et al. (1998), and Volynets et al. (1997). In column 4, calendar ages are given, calculated using earlier published individual radiocarbon dates, INTCAL98 calibration curve (Stuiver et al., 1998), and the OxCal v3.8 program (Bronk Ramsey, 1995, 2001). In field descriptions, "ash" designates grain size. Hb—hornblende; Ph—phlogopite. Question marks indicate estimated ages. In column 6, figures in parentheses show the maximum thickness found occasionally in a single section.

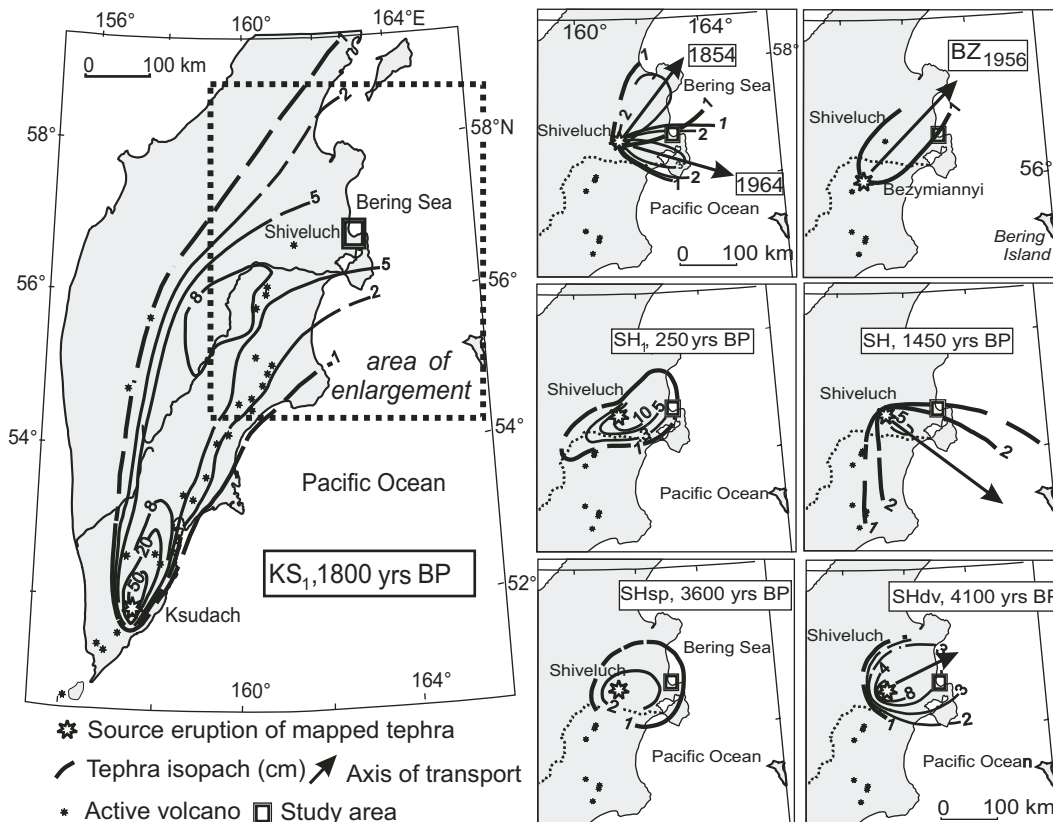


Figure 5. Isopach maps for major tephra layers deposited near study area over the last 6000 ¹⁴C yr (cf. Table 2). KS₁ revised from Braitseva et al. (1996); SH_{sp} revised from Volynets et al. (1997); other isopachs based on newly collected data (this study) and on Pevzner et al. (1998).

Shoreline History: Initial Stabilization, Subsequent Progradation, and Erosion

We infer that the Holocene sandy shoreline at Stolbovaia started forming earlier than the age of local basal peat above lagoonal sediments (Fig. 4, section 4). Basal peat from Stolbovaia was dated at 5380 ± 40 ^{14}C yr B.P. (Fig. 4; Fig. DR1 [see footnote 1]). Diatom analysis of the underlying deposits (Table DR2; see footnote 1) indicates a brackish water environment, which suggests that at ca. 5500 ^{14}C yr B.P. the lagoon was already separated from the Bering Sea. This basal-peat age is in accord with dates of 5500–6000 ^{14}C yr B.P. for the lowermost peat in the southern part of the lowland, at Chernyi

Yar (Figs. 2 and 4; Pevzner et al., 1998). In interior lowland sites (Fig. 2), basal peat overlying lake deposits is ~1000 yr younger (Fig. 4; Khalnitsa, Kultuk sections). We interpret these younger dates to indicate landward progradation of peat into a broad lagoonal lake.

Since establishment of the Stolbovaia shoreline ~5000–6000 yr ago, there has been ~100–300 m of net shoreline progradation along this coast, based on profile widths, but with erosional phases. Shoreline stabilization is dated by the oldest well-studied and -dated tephra in beach-ridge excavations, SH_{dv}, dated to 4100 ^{14}C yr B.P., and some older nonmarker tephras below it (Fig. 4, section 4). Although the Stolbovaia shoreline has prograded at times (see

profiles), it is not currently prograding except near river mouths, and most profiles have undergone recent erosion or stability. In a prograding system, the youngest beach ridges would contain only young tephras. However, at Stolbovaia, excavations within 25 m of the active beach in all profiles except 8 and 5—which are near river mouths—contain tephras at least 1800 yr old, and in at least half the cases, tephras at least 3500 yr old. Evidence for erosion includes the presence of alluvial sediment beneath young beach sediment in excavation 121 on profile 7 (Fig. 9; Fig. DR7 [see footnote 1]), as well as the proximity of peat and lagoonal sediments to the active beach in profiles 7, J4, and J3 (Fig. 9; Figs. DR12 and DR13 [see footnote 1]).

We roughly estimate the rate of erosion in the region of profile 8 (Fig. 9) by the maximum age of tsunami layers. The average inundation of tsunamis in this area, based on tsunami deposits younger than SH₁ (ca. 1600 A.D.), is ~150–200 m inland. Hence, the absence in this part of the coast of most tsunami layers older than SH₁₄₅₀ ash likely indicates that at this time (ca. 600 A.D.) the present coastline was at least 150–200 m farther seaward. This portion of the coast appears, then, to have been eroded for a minimum of 150 m between ca. 600 and 1600 A.D.; on that basis, the average erosion rate was 0.15 m/yr. Based on an extrapolation of erosion rates, we suggest that the

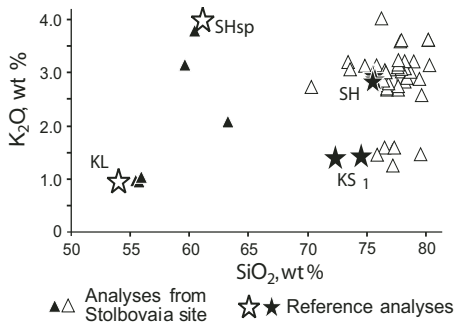


Figure 6. Chemical analyses of volcanic glass from tephra in the study area (Fig. DR1 [see footnote 1] for sample depths and descriptions; Table DR1 [see footnote 1] for chemical analyses). Dark triangles—dark tephra; white triangles—pumiceous tephra. Reference data for SH_{sp}, Volynets et al. (1997); Kliuchevskoi (KL), Kersting and Arculus (1994); KS₁, Volynets et al. (1999) and Rourke (2000); Shiveluch tephra (other than SH_{sp}), Rourke (2000).

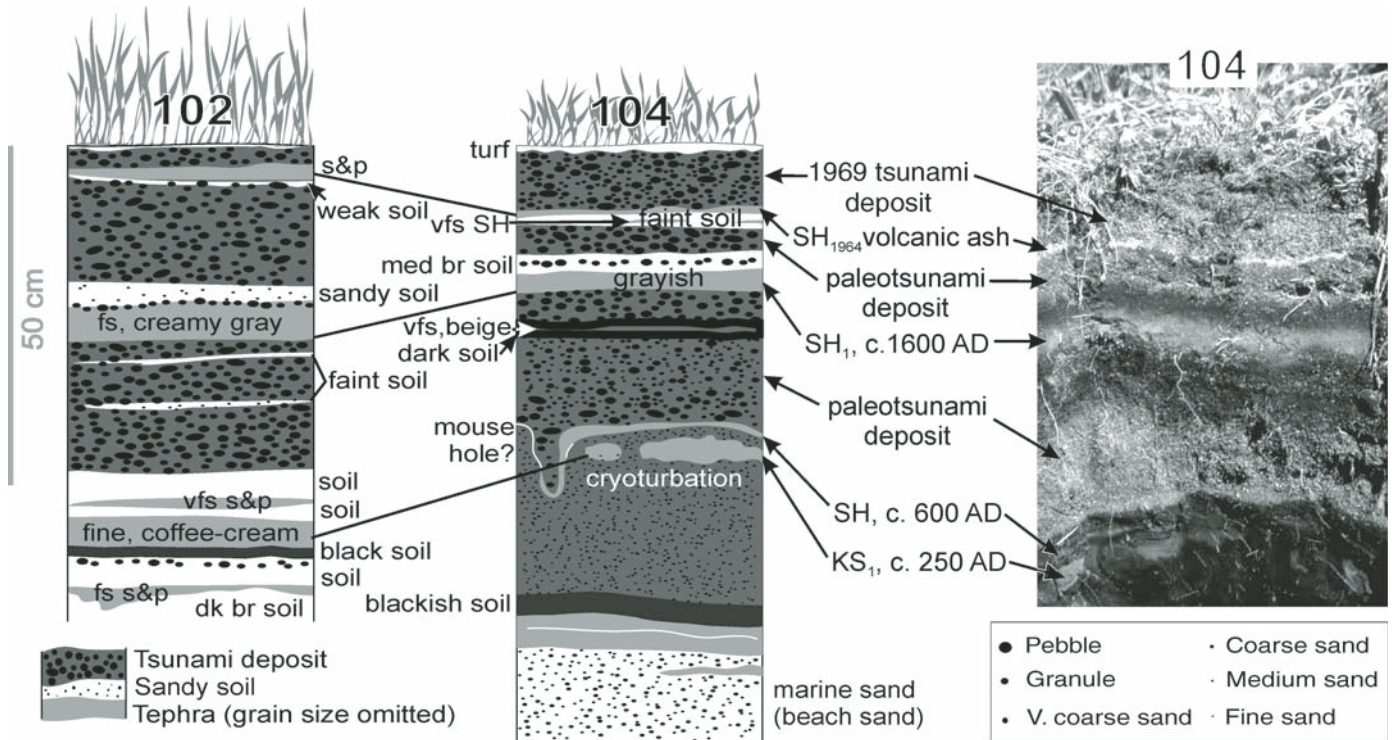


Figure 7. Photo (right) and redrawn field sketches (left) from profile 1 (excavations 102 and 104; Fig. 8; located in Fig. 3). Wording is principally from field descriptions: vfs—very fine sand; fs—fine sand; s&p—salt and pepper; for tephra designations (SH, KS), see Table 2.

indentation of the coast between profiles 7 and 8 (Fig. 3) indicates an erosion of ~0.5 km of shoreline over the past 3000 yr.

Evidence for Vertical Change

Most of the Stolbovaia coast has been tectonically stable in the Holocene, except for local areas where faults have offset the lowland. Net stasis (no vertical change) is interpreted in profile excavations where the elevation of the oldest soil is similar on the same profile to the lowest elevation of modern vegetation—i.e., the elevation at which soil begins to form.

There is facies evidence locally of vertical offset, between the northernmost profiles (7, J4, J3) and profile 8 (Fig. 9), where the offshore Pokatyi Canyon fault zone would project onshore toward the Grechishkin or Ust-Kamchatskii fault zone (Fig. 2). However, the connection between offshore and onshore faults has not been established (A. Kozhurin, 2005, personal commun.). On the basis of seismic stratigraphy, the net motion on the Pokatyi Canyon fault zone offshore is down to the southeast (Geist and Scholl, 1994), and our excavations confirm this motion at the shoreline. For example, in profile 7, north of the fault zone, the base of the peat is 3 m above

sea level, whereas in profile 8, south of the fault zone, the top of the peat section is only ~2 m above sea level and the base of the peat below present sea level (Fig. 9). The peat in profile 8 has been inundated by sand layers (tsunami or storm deposits) only since ~1100 ¹⁴C yr ago (section 5, Fig. 4). We interpret this appearance of sand layers to indicate that a few meters of subsidence occurred on the southeast side of the fault at that time or somewhat earlier, with sea-level rise and coastal erosion leading to greater susceptibility to tsunamis.

On the northwest side of the fault zone, lithologic changes and plant macrofossils in

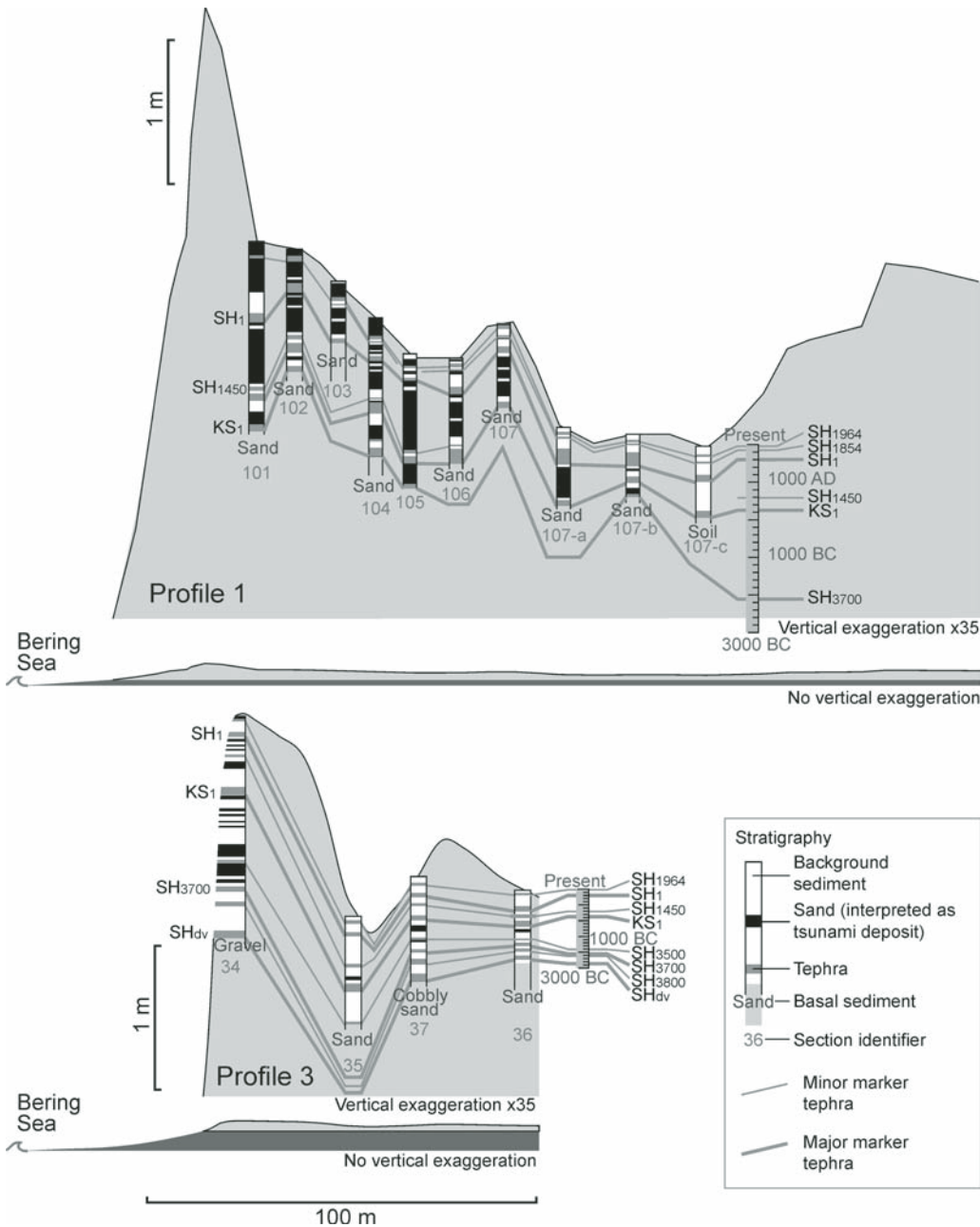


Figure 8. Examples of two profiles (located in Fig. 3) in form used for compiling tephra and tsunami data. Horizontal scale is constant. Part of the profile with excavations is shown at 35× vertical exaggeration, with sections at the same vertical scale; full profiles with no vertical exaggeration are shown beneath. Marker tephra (Table 2) are plotted to the right on a time scale, then correlated across the profile. “Background [non-event] sediment”—soil (usually) or peat. All profiles (Fig. 3) are in the GSA Data Repository (Figs. DR2–DR14).

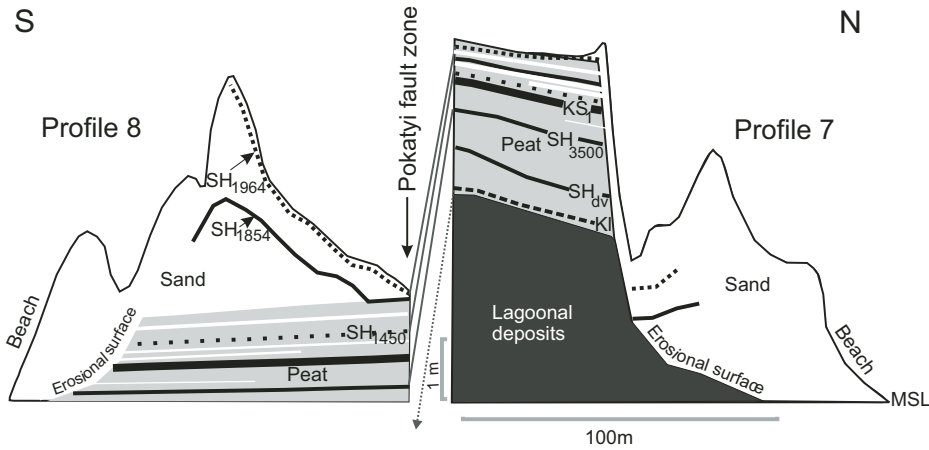


Figure 9. Interpreted profiles 7 and 8, on either side of the Pokatyi Canyon fault zone (Fig. 2; profiles DR7 and DR8; see footnote 1). To illustrate correlation and offset, one profile is flipped (beach in opposite direction). Waves have eroded peat and lagoonal sediments on both profiles, suggesting subsidence on both sides of the fault zone. The profile 8 section was expanded with sand (storm or tsunami deposits) some time between deposition of SH₃₅₀₀ and KS₁ (ca. 1800 B.C. and ca. 250 A.D.), suggesting significantly more subsidence south of the fault (see text discussion). At present uplift (since at least SH₁₈₅₄), neither profile is being eroded, suggesting either less storminess or uplift.

Carex-rich peat (on profile 7, Fig. 9; Podgorina 4, Fig. 4; Table DR2 [see footnote 1]) suggest both subsidence and uplift in the past 6000 yr. We estimate net uplift on the order of 1 m on the basis of the elevation of the youngest brackish-water lagoonal sediments ~1 m above modern high tide (profile 7, Fig. 9). Macrofossil changes in peat composition (Table DR2; see footnote 1), primarily of *Carex* species, as well as changes in the amount of mud (silt and clay) in the peat (Podgorina 4, Fig. 4), may represent episodes or trends in subsidence and uplift. However, because there has been no detailed study of the plants in this region, the following interpretations are tentative. The basal transition from brackish lagoonal sediments to peat is sharp and may indicate an uplift event. We interpret a subsequent transition from clean peat to muddy peat, with an attendant *Carex* species change, above the KL ash (ca. 6000 yr B.P.), to suggest subsidence. We interpret the highest sample analyzed, distinctly different in *Carex* species from 3000 yr worth of underlying peat, and containing woody shrubs, to indicate uplift sometime between deposition of that sample (just under SH₁) and the underlying sample (below SH₁₄₅₀). Because we did not sample continuously, and age control is based on tephra, we cannot narrow down the age of this transition more than between deposition of these two tephra, or 1400–400 yr ago. Uplift in this time period could coincide with aforementioned subsidence on the south side of the fault.

Tsunami Deposits at Stolbovaia—DATA AND INTERPRETATION

The source for tsunami deposits at Stolbovaia is primarily beach sediment—moderately well rounded, medium to coarse sand and fine gravel—deposited across vegetated surfaces. Some tsunamis may have traversed a snow-covered (rather than vegetated) surface. For example, Minoura et al. (1996) describe an April 1923 tsunami deposit across snow, south of Kamchatskii Peninsula (Fig. 1B). However, Kamchatka beaches are frozen from winter to early spring, and southern Ozernoi Bay is typically filled with sea ice in the winter, so tsunami deposits and tsunamis (as in Otsuka et al., 2005) are suppressed during this time.

Sand deposits interbedded with soil, peat, and tephra from 53 sections in 13 profiles are summarized in Figure 10, which also provides the modern elevation of excavations and their distance from the shoreline. We interpret these sand layers to be tsunami deposits, with possible exceptions, as discussed later. In a prior section of this article, we reviewed our criteria for tsunami-deposit identification and noted that we use the 1969 Ozernoi tsunami as a guide.

We use the landward extent and elevation of interpreted tsunami deposits as indicators of tsunami size; deposit thickness and grain size can also be indicators of tsunami size but are less reliable because these characteristics can be strongly controlled by local effects. Tsunami size

is formally characterized by *runup*, the elevation of the tsunami at maximum penetration distance, and *inundation*, the maximum penetration distance, measured in the direction of tsunami flow, typically perpendicular to the shoreline. On a stable shoreline, we can take the current elevation and extent of a tsunami deposit as an indicator of runup and inundation, but on unstable shorelines we must take into account changes in relative sea level and in shoreline location. Even for young tsunami deposits, runup estimates may be inaccurate if there has been uplift or subsidence associated with recent earthquakes. Moreover, inundation distances based on tsunami deposits are minima, because tsunamis can penetrate farther than their deposits (e.g., Nishimura and Naomichi, 1995; Hemphill-Haley, 1996; Higman and Bourgeois, 2002).

Tsunamis Since ca. 1600 A.D. and Their Deposits

1969 Ozernoi Earthquake and Tsunami

On 23 November 1969, a Mercalli intensity 7–8 earthquake of moment magnitude 7.7 jolted the Ozernoi Peninsula (Fig. 1C). Fedotov and Gusev (1973) interpreted this earthquake as strike-slip, but Cormier (1975) reinterpreted it as a thrust fault on the basis of data from global seismograph networks. Using body waveform analysis, Daughton (1990) also found a thrust-fault-plane solution, striking N50°–80°E and dipping 5°–10°N. The thrust-fault solution is consistent with the tsunami generated by this deformation, and this solution is also an important component of our overall tectonic interpretation.

The 1969 Ozernoi earthquake was followed by a tsunami with reported tsunami heights of 5–7 m from Karaginskii Bay south all the way around the Ozernoi Peninsula; runup locally as high as 10–15 m was reported (Table 1; Zayakin, 1981). North to Lavrov Bay and southeast to Bering Island, reported heights were 1–3 m, and the tsunami was recorded on Kamchatka tide gages to the south (Zayakin and Luchina, 1987). There are no published tsunami reports from southern Ozernoi Bay, although there was a small military outpost there at the time. In a 1999 unprompted account, a local hunter said the military post was abandoned after a large “storm” in 1969 destroyed one house and damaged others; we think this “storm” was the 1969 tsunami.

Deposits from the 1969 Ozernoi Tsunami and “1999 Surge”

The layer we interpret to be the 1969 Ozernoi tsunami deposit (Fig. 7), present in 34 of 58 sections (Fig. 10), is a coarse-grained sand (to fine gravel), typically 5–20 cm thick (varying from a trace to >50 cm) almost directly

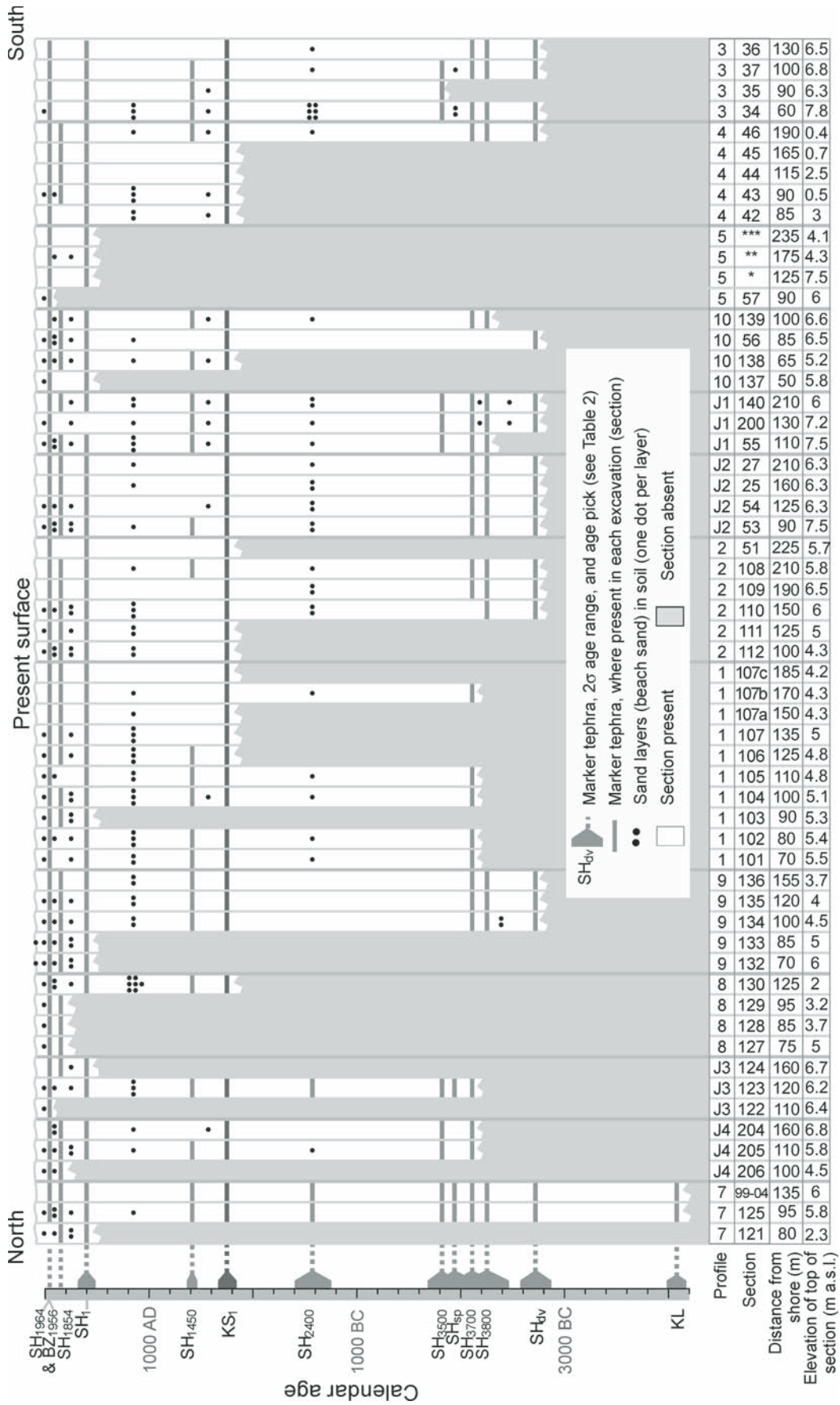


Figure 10. Summary of tephra and marine-sand deposits (in most cases interpreted to be tsunami deposits; see text) from all sections. Tephras are plotted at a given time line, which is our age pick (selected age) for each marker tephra, with the 2 sigma (2σ) range of ^{14}C calendar ages (Table 2) shown in the left column. SH_{2400} has not been correlated regionally and may be older; it was not used in frequency calculation, but it is used for local correlation. Profiles are arranged from north to south (left to right), and within each profile, sections are shown from seaward to landward (left to right). All profiles are in the GSA Data Repository (Figs. DR2–DR14); m a.s.l.—meters above sea level.

overlying SH₁₉₆₄ tephra, and overlain by well-developed turf (Fig. 7). The deposit is massive, with a mixed grain-size distribution ranging from medium sand to small pebbles; the median is typically very coarse sand and granules. This deposit extends ~100–150 m from the shoreline, overtops beach ridges at least 6–8 m high, and persists landward at elevations of 4–7 m. Because the deposit overlies SH₁₉₆₄ tephra (deposited in November 1964), it is younger than the Kamchatka 1952, Chile 1960, and Alaska 1964 (March) earthquakes and attendant tsunamis. None of these large teletsunamis had local heights in the Ozernoi region comparable to the 1969 Ozernoi tsunami (Table 1).

Between the 1998 and 1999 summer field seasons, an event we call the “1999 surge” (it could have been autumn 1998) left a deposit in the most seaward two sections (132, 133; Fig. 10) on profile 9 (Fig. DR9; see footnote 1), but nowhere else (Fig. 10). A wave or waves from this surge washed over the 6-m-high beach ridge at profile 9 and ran down the back side for ~20 m. On the beach near profile 9, between the 1998 and 1999 field seasons, surface features on the backbeach were washed away. This “1999 surge” also affected the backbeach on other profiles (1–5, 8) but did not wash over the vegetated beach ridge. There are no local meteorological observations from 1998 to 1999, so this deposit could be from a storm or from a small landslide-generated tsunami. The deposit does not meet our criteria for tsunami-deposit identification.

A comparison of the “1999 surge” and the 1969 tsunami deposits can offer a guide for interpreting older tsunami deposits at Stolbovaia. Compared to the “1999 surge” deposit, the 1969 tsunami-deposit thickness, massive structure, and extent require flooding of the entire surface rather than local wave washover. The thickness ratios between the 1969 tsunami and the “1999 surge” deposits in sections 9–132 and 9–133 are 25:1 and 16:1, and the 1969 tsunami deposit at profile 9 extends ~50 m farther inland than the surge deposit. Despite these apparent differences, and the lack of the “1999 surge” deposit in any other excavations, we use this “1999 surge” deposit as a caution. Thus, we exclude most proximal excavations (<~50 m landward of the current beach-ridge crest) in our analysis of tsunami-deposit recurrence.

Other Young Tsunami Deposits at Stolbovaia (since SH₁, ca. 1600 A.D.)

Four tsunami deposits lie below SH₁₉₆₄ and above SH₁, although not every deposit is in every section (Fig. 10). We reject historical (tele)tsunamis as sources for these deposits and interpret them all to be from locally generated tsunamis in the southwestern Bering Sea.

Two of these deposits are between SH₁₉₆₄ and SH₁₈₅₄, but we cannot attribute either to a historical tsunami (Table 1). Neither is as extensive or as thick as the 1969 tsunami deposit; however, each reaches inundation distances at least 50 m from the beach crest and well beyond the “1999 surge” deposit. In the few cases in northern profiles where BZ₁₉₅₆ tephra is also present (Fig. 10), no tsunami deposit lies between it and SH₁₉₆₄. However, in southern profiles, a tsunami deposit is close to and below SH₁₉₆₄. Candidate historical tsunamis for the production of the two tsunami deposits between 1964 and 1854 are Alaska 1964, Chile 1960, Kamchatka 1952, and Kamchatskii Bay 1923 (April); but we reject these sources because none of their tsunamis produced tsunami heights on Bering Island >4 m (Table 1), whereas the deposits in Stolbovaia, farther away and around the corner from Bering Island, require a runup of 5–6 m.

The deposit close to and below SH₁₉₆₄ may have been generated by an unrecorded tsunami from the local 15 April 1945 earthquake (57°N, 164°E; magnitude ~7 in Kondorskaya and Shebalin, 1982), or possibly from a submarine landslide it triggered. The instrumental seismic record from 1945 isn’t good enough to determine a precise location, source mechanism, or moment magnitude for this earthquake; Gusev and Shumilina (2004) estimate an M_w of 7.3. The main shock produced two aftershocks greater than magnitude 6. There is no known written record of a 1945 tsunami on the Bering Sea coast of Kamchatka.

Two tsunami deposits lie between SH₁₈₅₄ and SH₁ (ca. 1600 A.D.). At least one of these is thicker and more extensive than the 1969 deposit. The historical catalogue in this time period is less complete than for the twentieth century, but the 1737 south Kamchatka tsunami may have deposited sand at Stolbovaia. This tsunami is reputed to have run up 30 m or more on Bering Island on the basis of an observation in 1741 by Steller of sea-mammal bones at a high elevation (Zayakin and Luchinina, 1987), although this account and interpretation are questionable. If the 1737 earthquake and tsunami were comparable to Kamchatka 1952 (Table 1; Gusev and Shumilina, 2004), then the 1737 tsunami would not have left a deposit at Stolbovaia.

Attribution of post-1600 Tsunami Deposits to Ozernoi-like Earthquakes

All four of these deposits require a runup of 4–6 m in south Ozernoi Bay, with minimum inundation distances of 50–100 m, and with the possible exception of local submarine landsliding, we argue that such a runup and inundation require local, Ozernoi-scale earthquakes. The specific fault responsible for the 1969 Ozernoi and predecessor earthquakes is still not

mapped, but a possibility is that the boundary of the Olyutorsky terrane (Garver et al., 2000; Fig. 1) has been reactivated. Another possible tsunami source at Stolbovaia is local slip along the Pokatyi fault zone, with accompanying landslides in its submarine canyon (Fig. 2).

As noted previously, whereas the 1969 Ozernoi produced a deposit along the Stolbovaia field site, the largest tsunamis of the twentieth century (Table 1) did not generate sufficient runup at Stolbovaia to do so. No other tsunamigenic earthquake in the historical catalogue, with the possible exception of the 1737 south Kamchatka, produced greater runup on Bering Island or points north than the 1969 Ozernoi. Although one could argue that the historical record is geographically spotty, we have studied several other sites north of the Commander Islands (e.g., Soldatskaia Bay, Ozernoi Cape; Figs. 1 and 2) that exhibit similar 1969-tsunami deposits and similar tsunami frequencies (Fig. 11). We interpret this regional distribution of deposits as evidence that most of them were generated by tsunamis from earthquakes in the southwestern Bering Sea, similar to the one in 1969.

Other possible sources for these deposits, specifically storms, or tsunamis generated by non-seismic submarine landslides, are less likely. The Stolbovaia site lies at the head of a submarine canyon (Fig. 2). This canyon has an active normal (or dip-slip) fault on its northern margin, so earthquake-triggered submarine landslides could have generated some tsunamis. But this canyon is limited in extent; and as can be seen on air photos, currently most sediment from the Stolbovaia River drainage is trapped inland so that the canyon is not receiving much sediment. Thus, landsliding should not generate large tsunamis. Moreover, nearby sites that are neither along a fault zone nor near a submarine canyon, over a latitude range of more than 1°, have similar recurrence rates of tsunami deposits (Fig. 11; based on our field data). As with 1969 Ozernoi, such a pattern argues for regional tectonic sources rather than local landsliding to explain the record.

Millennial-Scale Record—Estimating Tsunami Ages and Recurrence Rates

Factors and Caveats

Excavations at the Stolbovaia site permit us to examine tsunami-deposit frequency back to the SH_{dv} tephra layer, or ~4500–4800 calibrated yr ago (Table 2), and to compare these frequencies to other sites, with the following caveats. First, we should expect geographic variance in tsunami deposits owing to different earthquake source characteristics and also to bathymetric effects on tsunami propagation. Second, preservation of tsunami deposits and

tephras is variable and usually best in sections with rapid accumulation rates and generally better in younger deposits. Third, with more excavations, it is likely that more tsunami deposits will be identified. At Stolbovaia, for example, the number of excavations that expose a given time interval decrease with increasing age (see Fig. 10), so fewer tsunami deposits may be recognized in older parts of the record. Fourth, frequency statistics are affected by modern and paleocoastal morphology and local topography: some excavations are as low as 3–5 m above sea level, and most are >5 m above sea level. Multiple excavations along a profile can alleviate this problem. Finally, the history of coastal accretion, erosion, and relative sea-level change must be addressed.

Effects of erosion. The sites that Stolbovaia is compared to in this discussion (Fig. 11)—Ozernoi Cape (Martin et al., 2004) and particularly Soldatskaia Bay (Kravchunovskaya et al., 2004)—have prograding shorelines, as shown by progressively disappearing older tephras in shoreward beach ridges. As discussed previously, at Stolbovaia many profiles show that early progradation was followed by late Holocene coastal erosion. Therefore, in older parts of the section, because the shore-proximal region has been eroded away, only deposits from larger tsunamis will still be preserved. In a prior discussion, we attempted to quantify an erosion rate near profile 8 (Fig. 9) on the basis of tsunami deposits.

Outlier excavations. A quandary in calculating tsunami-deposit frequency is whether or not to use single excavations that exhibit more sand layers (between certain marker tephras) than do other excavations. Are such cases exceptionally well-preserved tsunami records, or records of storms as well as tsunamis? At Stolbovaia, sections 130 and 34 (Fig. 10) fit this category. In section 130, we interpret the sand layers between SH_1 and SH_{1450} as tsunami deposits rather than storm deposits, because they are interbedded with fresh-water peat and because the coast has been eroding (Fig. 9); so when these sands were deposited, the site was ~50–200 m farther from the shoreline. Section 34 on profile 3 (Fig. 8) is an eroded beach cliff; at 7 m elevation, it exhibits neither the 1969 tsunami deposit nor the “1999 surge” deposit. There is no evidence for sea-level change during the time represented at this site; hence, we interpret the sand layers (between KS_1 and SH_{3500}) in section 34 to be deposits from tsunamis larger than the 1969 tsunami.

The Record at Stolbovaia and Neighboring Bering Sea Sites

Subject to the caveats outlined previously, we extend our paleo-tsunami analysis back several millennia using multiple excavations, good age

control, and a historical tsunami for calibration (Figs. 10 and 11). Marker tephras are key to our analysis of tsunami frequency. More important than the precise ages of these tephras is the fact that they are time lines (isochrons). Most of our sections span at least the past 2000 yr (back to KS_1), and many span ~4500 yr. We calculated the total number of tsunamis from the maximum number of deposits recorded within each tephra-delimited time interval (Fig. 11).

Over the past 2000 yr, three sites over ~1° latitude in the southwestern Bering Sea have large-tsunami (>5 m runup; >50 m across a vegetated surface) frequencies of >5 per thousand years, or an average recurrence of at least one large tsunami per 200 yr. Figure 11 illustrates summary tsunami-deposit frequencies for these three sites—Stolbovaia, Ozernoi Peninsula to the north, and Soldatskaia Bay to the south (Figs. 1 and 2). We show both the frequency calculated between each marker tephra and the average frequency, using only widely separated (temporally) and widespread marker tephras. The former method produces greater variability, because one tsunami deposit between two closely spaced tephras will give a frequency that may average out over centuries to millennia.

At all three sites, tsunami-deposit frequency decreases in older sections, likely owing largely to poorer preservation and to fewer observations in older deposits. If we are correct in our argument that the Stolbovaia coastline has undergone net erosion, then the older record we examined should contain fewer tsunami deposits in any case. Thus, only deposits from the largest tsunamis, with the greatest inundation distances, should be preserved in the older section at Stolbovaia. This argument is bolstered by data from Soldatskaia (Fig. 11; Table DR3 [see footnote 1]; Kravchunovskaya et al., 2004), which has been prograding since SH_{4700} (ca. 4700 yr B.P.) and shows more uniform frequencies back through time (Fig. 11).

DISCUSSION AND CONCLUSIONS

Although the Stolbovaia site is not along an active subduction zone, at this site we have documented at least 12–15 tsunami deposits in ~4500 yr, a recurrence comparable to active subduction zones around the North Pacific. Tsunami-deposit recurrence at this and other sites along the southwestern Bering Sea is about half the recurrence along the central coast of Kamchatka (Kronotskii Bay, Fig. 1; Zhupanova site of Pinegina et al., 2003), a seismically active subduction zone. The recurrence at Stolbovaia is comparable to the 20 deposits in 9000 yr documented on Hokkaido, at the southern end of the Kuril-Kamchatka subduction zone (Nanayama et al., 2003). Recur-

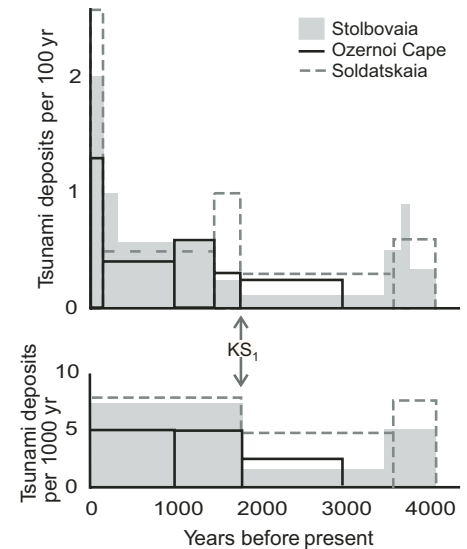


Figure 11. Tsunami-deposit frequency calculated for Stolbovaia (this paper), and estimated for Ozernoi Cape (located in Fig. 1B) and Soldatskaia (located in Fig. 2). Preliminary data summary for the latter two sites are in Table DR3 (see footnote 1). The upper diagram shows deposit frequency (calculated per hundred years) between every marker tephra present at each site. The lower diagram shows deposit frequency (calculated per thousand years) using only the most widespread marker tephras. Data are less complete from parts of sections older than KS_1 (Fig. 10).

rence is greater than along the Alaska and Cascadia subduction zones, where the paleoseismic and tsunami-deposit records indicate great ($M_w > 8$) subduction-zone earthquakes on an average of once every 500 yr (Saltonstall and Carver, 2002; Hamilton et al., 2005; Atwater and Hemphill-Haley, 1997; Kelsey et al., 2005).

For Stolbovaia and other southwestern Bering Sea sites, we have evaluated, and consider unimportant, sources for tsunamis other than local earthquakes such as 1969 Ozernoi, which we have used as a comparative scale. Other than 1969, there are no known historical tsunamis in the southwestern Bering Sea with runup of 5 m or more. We have shown that these Bering Sea sites are not subject to large teletsunamis and have argued that the tsunami-deposit record at Stolbovaia isn't just from submarine landslides in the Pokatyi submarine canyon because of the regional rather than local distribution of tsunami deposits. Finally, when we compare tsunami-deposit frequency in the southwestern Bering Sea to other localities worldwide, those sites face many of the same challenges of interpretation.

Our study of tsunami-deposit recurrence at Stolbovaia is important for its natural-hazards implications. Although previous studies have considered tsunamis in the Russian Bering Sea (Gusiakov and Marchuk, 1997; Melekestsev and Kurbatov, 1998; plus tsunami catalogues), our study is the first to document a long-term history of Bering Sea tsunami deposits sufficient to consider recurrence intervals. Using the large 1969 Ozernoi tsunami as a benchmark (>5 m runup along >1° latitude of coastline), we have documented comparable or larger tsunamis recurring on average every 200 yr (Fig. 11). Although this coastline is not heavily populated, such a recurrence interval requires natural-hazards consideration—e.g., for local fishing villages and fleets, meteorological stations and lighthouses, and military outposts. On the basis of the regional setting and the 1969 Ozernoi event, tsunamis generated in this region will not produce damaging teletsunamis, unlike subduction zones such as those noted previously.

The history of seismicity (Fig. 1C) and paleoseismicity (this paper) along the coast of the southwestern Bering Sea, including the M_w 7.7 Ozernoi 1969 tsunamigenic earthquake, indicates that the area is actively undergoing compression and that the traditional two-plate model cannot explain this compression. Historical seismic activity along the Aleutian shear zone dies off north of the Komandorskii Island block (KIB) (Fig. 1), so shear from the Pacific plate does not appear to be impinging on the Stolbovaia region. Moreover, current geophysical and petrologic models of the Pacific plate at its northwest corner support the idea that it should not be transferring strain as far north as Stolbovaia. These “torn-slab” models show that the northwestern Pacific plate has torn away and dropped off deeper into the mantle in this region (e.g., Yagodinski et al., 2001; Levin et al., 2002; Park et al., 2002; Portnyagin et al., 2005), so the Pacific plate would not be overriding the southwestern Bering Sea coast.

Our documented record of tsunamis at Stolbovaia demands that we ask, “What is the source of tectonic activity along the southwestern Bering Sea?” We favor a multiplate model with the Bering and Okhotsk blocks converging in this region (Fig. 1A). If shear is *not* being transmitted to the region from the Pacific plate, then compression in the southwestern Bering Sea *must* be explained by some plate (or block) motion other than convergence of the Pacific and North American plates. Either eastward movement of the Okhotsk block, or westward movement of the Bering block, or both, can explain the historic and paleoseismic record. This conclusion is consistent with other analyses of motion of the Okhotsk and Bering blocks (Riegel et al.,

1993; Mackey et al., 1997; Mackey et al., 2004) on the basis of seismic records (e.g., Fig. 1C).

The multiplate model (Fig. 1A), including the seismic and tsunami-deposit evidence that there is active compression along the southwestern Bering Sea coast, leads to some predictions. We would expect to find evidence of deformation such as uplifted and tilted marine terraces along these shorelines; indeed, such terraces have been mapped on Ozernoi Peninsula (Pedoja et al., 2004), and we have observed them in photos and digital elevation maps of Karaginskii Island. There should be other geological and geomorphological evidence of these plate boundaries, such as fault lineaments found in parts of northeastern Siberia (Mackey et al., 2004); these features have not been identified on Kamchatka. The precise location of the boundaries of these proposed blocks, and their behavior as rigid and independent plates (or not), are still unsolved problems that require further examination.

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