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Key Points:

- Tectonic subsidence produces variable receiving basin depth, and drives lobe-scale avulsions by modifying delta-topset gradient
- Channel-scale avulsions occur during periods of tectonic quiescence, and disperse sediment to nourish the deltaic shoreline
- Hierarchical avulsion styles could lead to preservation of discrete stratal packages that contain predominately deep channels

Supporting Information:

Supporting Information may be found in the online version of this article.

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Impacts of Tectonic Subsidence on Basin Depth and Delta Lobe Building

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Abstract Channel avulsions on river deltas are the primary means to distribute sediment and build land at the coastline. Many studies have detailed how avulsions generate delta lobes, whereby multiple lobes amalgamate to form a fan-shaped deposit. These studies often assume a steady subsidence and uniform basin depth. In nature, however, lobe building is disrupted by variable subsidence, and progradation of lobes into basins with variable depth; conditions that are prevalent for tectonically active areas. Herein, we explore sediment dispersal and deposition patterns across scales using measurements of delta and basin morphology compiled from field surveys and remote sensing, collected over 150 years, from the Selenga Delta (Baikal Rift Zone), Russia. Tectonic subsidence events, associated with earthquakes on normal faults crossing the delta, displace portions of the topset several meters below mean lake level. This allogenic process increases regional river gradient and triggers lobe-switching avulsions. The timescale for these episodes is shorter than the predicted autogenic lobe avulsion timescale. During quiescent periods between subsidence events, channel-scale avulsions occur relatively frequently because of in-channel sediment aggradation, dispersing sediment to regional lows of the delta. Avulsion settings for the Selenga Delta preserve discrete stratal packages that could contain predominately deep channels. Exploring the interplay between tectonic subsidence and sediment accumulation patterns will improve interpretations of stratigraphy from active margins and basin models.

Plain Language Summary River deltas distribute sediment and build land in coastal regions by abruptly shifting course through a process called channel avulsion. The fan-shaped morphology of river deltas arises from multiple avulsion events that deposits sediment in a radial fashion over time. Our understanding of how deltas build such morphology assumes that size of the downstream basin, such as a lake or ocean, is constant over time. However, geological activities like earthquakes and basin subsidence alter the size and shape of the delta. We complied and analyzed 150 years of delta morphology data from the Selenga Delta (residing in a tectonically active area) in Russia to understand how earthquakes and subsidence impacts channel avulsions and delta landforms. We determined two distinct avulsion styles for the Selenga Delta: a regional-scale avulsion that is driven by earthquakes, and a local-scale avulsion that is caused by sediment deposition. The two distinct scales of avulsions blend over time to shape the delta system. In addition, the regional-scale avulsion produce unique subsurface records that can be used to understand the history of a delta. Our work highlights the importance of understanding the variety in downstream basin processes that impact delta morphology.

1. Introduction

River deltas prograde basinward by distributing sediment over the topset, foreset, and bottomset. A major contributor to spatiotemporal variability in sediment dispersal are channel avulsions, which relocate delta lobes and depocenters (Chadwick et al., 2019; W. Kim et al., 2010; Reitz & Jerolmack, 2012; Swenson, 2005). With multiple avulsions, delta lobes amalgamate to produce a fan shape that continues to be nourished by the distributary channel network (B. Carlson et al., 2018; Ganti et al., 2014; Moodie et al., 2019; Piliouras et al., 2017). Theoretical and experimental evidence suggests that, over time, delta lobe growth reaches a state of sediment-transport

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equilibrium, known as alluvial grade, characterized by sediment bypass of the topset with delivery to the foreset (B. Carlson et al., 2018; Y. Kim et al., 2013; Muto et al., 2016; Posamentier & Allen, 1999; Richards et al., 1998). Alluvial grade and channel avulsions are impacted by autogenic and allogenic processes that alter upstream and downstream boundary conditions, thereby affecting delta steady-state dynamics (Wang et al., 2019). Constraining the interplay of these processes over a range of timescales is thus critical to improving delta-evolution models. Such scientific developments are useful in various modern settings to combat land loss, as well as to evaluate stratigraphy in ancient settings (B. N. Carlson et al., 2021; W. Kim et al., 2006, 2009; Straub et al., 2009; Syvitski et al., 2009).

Alluvial grade of a delta lobe can be assessed using the grade index (G_{index} ; Muto et al., 2016):

$$G_{\text{index}} = \frac{1}{1 + 2h_* + \alpha_* h_*^2},$$

$$\alpha_* = \frac{S_{\text{fan}}}{S_{\text{basin}}},$$

$$h_* = \frac{H_{\text{basin}}}{\bar{R}S_{\text{fan}}},$$
(1)

where α_* and h_* are normalized delta topset slope and basin water depth, respectively, S_{fan} is topset slope, S_{basin} is basin slope, H_{basin} is basin depth, and \bar{R} is the mean delta radius. Herein, $G_{\text{index}} \rightarrow 0$ indicates a river delta which achieved alluvial grade and $G_{\text{index}} \rightarrow 1$ indicates sediment imbalance. Since most delta systems have relatively low topset gradients and flow depth, basin depth (H_{basin}) is one of the most important parameters that impacts alluvial grade. This variable is often affected by tectonic subsidence (B. Carlson et al., 2018). For example, deltas on active margins usually maintain deep basin depth, and therefore achieve alluvial grade, whereby aggradation on the topset is negligible and distributary channels are immobile and possess well-developed levees (Muto et al., 2016; Wang et al., 2019). While accommodating sediment dispersal to the foreset, a deep receiving basin depth limits shoreline progradation because it takes longer to fill the space at the delta front (B. Carlson et al., 2018).

Alluvial grade also affects the development of stratigraphy. Specifically, stratigraphic completeness, that is, the preservation of genetically related fluvial-deltaic facies from proximal topset to distal foreset, is viewed as a competition between accommodation and sediment supply (Straub et al., 2013). Deltas at alluvial grade may preferentially preserve strata in the foreset due to limited topset aggradation (Y. Kim et al., 2013). Stratigraphic completeness of delta deposits can be approximated by the filling index, *B* (Liang et al., 2016):

$$B = \frac{dV_{\text{accomm.}}/dt}{Q_{\text{supply}}},$$
(2)

where dV_{accomm}/dt is the change volume of accommodation, per-unit-time, generated by subsidence, and is closely associated with basin depth (H_{basin}). Q_{supply} is sediment supply. When B > 1, accommodation outpaces sediment supply, and delta progradation is limited; conversely, when B < 1, sediment supply outpaces accommodation, facilitating delta progradation (W. Kim et al., 2010; Kopp & Kim, 2015; Liang et al., 2016; Reitz et al., 2015; Straub et al., 2013).

Alluvial grade also affects the size of preserved sedimentary structures, such as lateral accretions produced by mobile channels. For example, immobile distributary channels at alluvial grade and preferential reworking of smaller bedforms preserve the largest dunes that develop during floods (Ganti et al., 2020; Wu et al., 2020). One way to quantify the preservation potential of different sedimentary structures is to use the preserved extremality index (Ω), a metric ranging from 0 to 1 (Ganti et al., 2020):

$$\Omega = \frac{100 - 2\tilde{p}}{100},\tag{3}$$

where \tilde{p} is the median percentile of the preserved topography (size of sedimentary structure). $\Omega \rightarrow 1$ indicates that large sedimentary structures deposited during low-frequency, high-magnitude events dominate preserved strata; conversely, $\Omega \rightarrow 0$ indicates that deposits formed during high-frequency, low-magnitude events, that is, "ordinary features," dominate preserved stratigraphy.

Two assumptions are often made about alluvial grade and development of deltaic stratigraphy: time-continuous subsidence and uniform receiving basin depth (e.g., Liang et al., 2016). In nature, however, basin geometry is





Figure 1. (a) Lake Baikal and the Selenga River delta, located in southeastern Siberia, Russia. (b) Contour map of Lake Baikal water depth, produced using data from De Batist et al. (2006). Hydrological data are collected from the seven main distributary channels are shown on the map. Bathymetric maps of (c) Cherkalovo and (d) Proval Bays, produced from data collected in 2001 by Vologina et al. (2007, 2010) and Pavlov et al. (2019).

modified by spatially variable subsidence and filling of accommodation. In tectonic settings, for example, multiple faults may be active, generating variable receiving basin depth (Dong et al., 2016; Martinsen & Bakken, 1990; C. Scholz et al., 1998; Shchetnikov et al., 2012; Vologina et al., 2010). Rift basins are well-documented sediment sinks, however, the impacts of tectonic subsidence on basin depth and delta lobe building remain elusive (Ravnås & Steel, 1998). Field evidence indicating how delta morphology and lobe growth are impacted by alluvial grade is also limited (Y. Kim et al., 2013; Ganti et al., 2014; Muto et al., 2016; Wang et al., 2019).

Herein, data from the Selenga River delta, Russia, are used to assess the effects of tectonic subsidence on basin depth and delta lobe building over 150 years. Specifically, existing theory for alluvial grade is applied to understand how tectonic subsidence modifies basin depth, delta topset morphology, shoreline position, sediment transport, and avulsion timescales. These findings are leveraged with literature-compiled subsurface evidence from the Selenga Delta to describe stratigraphic architecture and completeness about the Selenga system specifically, and inform about deltas in tectonically active areas broadly.

2. Lake Baikal and the Selenga River Delta

The Selenga River delta is located at the southeastern shore of Lake Baikal, Russia (Figure 1a; Colman, 1998; Il'icheva et al., 2015; C. Scholz et al., 1998). This basin is formed by rifting that was initiated ~35 million years ago (Krivonogov & Safonova, 2017; Logatchev, 1974, 2003; Mats & Yefimova, 2015; C. Scholz et al., 1998). Lake Baikal extends over 700 km in length and has an average width of 60 km, and a maximum depth of 1,650 m (C. Scholz et al., 1998). Lake Baikal's water level has remained relatively stable and the mean lake volume is interpreted to be roughly constant for over the past ~100 Kyr. (Colman, 1998; C. Scholz et al., 1998). Additionally, there is no evidence for major tectonism that would substantially modify the basin configuration and potentially impact the lake volume during the last 100 Kyr (Krivonogov & Safonova, 2017; Logachev, 2003).



The recent construction of the Irkutsk Hydroelectric Power Plant in 1960 has increased the lake level by ~ 1 m (Il'icheva et al., 2015). Seismic imaging indicates that sediment thickness is 4–5 km in the South Baikal Basin, and 7.5–10 km in the modern Selenga Delta front (Hutchinson et al., 1992). The variable thickness of sediment accumulation, and underlying bedrock highs and lows, have created a bathymetric saddle between the South Baikal Basin and the Central Baikal Basin, where the Selenga Delta is situated (Figure 1b; Hutchinson et al., 1992; C. A. Scholz & Hutchinson, 2000).

The delta channel network maintains variable bed and bank sediment size, vegetation, and morphology across the alluvial topset, extending 35 km from the apex, to the delta shoreline (Dong et al., 2016, 2019, 2020; Il'icheva, 2008; Il'icheva et al., 2015; Pietroń et al., 2018). Both median bed- and bank-sediment grain size fine downstream, from gravel and sand at the apex to silt and very-fine sand at the shoreline (Dong et al., 2016).

On timescales of 10^2-10^3 years, delta morphology is influenced by seismic events. Specifically, a portion of the subaerial delta subsides by up to 4 m (Lunina & Denisenko, 2020; Shchetnikov et al., 2012), a length that exceeds the mean distributary channel depth (2.7 m) of the delta (Dong et al., 2019). For example, in association with a recent (1862) seismic event (M 7.5), 200 km² of the delta downdropped by ~3 m, forming Proval Bay (Figure 1d; Lunina & Denisenko, 2020; Vologina et al., 2007, 2010). This subsidence event steepened the regional slope and drove a lobe avulsion that diverted water and sediment from the central region of the delta to fill the newly formed bay (Figures 1b and 1d).

Other embayments have been formed similarly, and are distributed around the delta, including Cherkalovo and Posolsky Bays (Figures 1b and 1c; Shchetnikov et al., 2012). Cherkalovo Bay was estimated to have formed between 1765 \pm 235 and 2905 \pm 205 years before present, based on ΔC_{14} dates from sediment cores (Figure 2a; Pavlov et al., 2019). Posolsky Bay, just south of the delta, formed ~500–600 years ago (Figure 2a; Shchetnikov et al., 2012). Based on these historical records, the recurrence interval of morphologically impactful earthquakes that creates embayments on the delta is 340–2,600 years (Table 1). We refer to this interval as the tectonic timescale (T_t) in discussions below.

According to the grade index formulation (Equation 1) proposed by Muto et al. (2016), the Selenga Delta is considered a "force grade" or "fixed" system (Figure 1c therein). Specifically, the delta is essentially "frozen" because accommodation volume at its downstream boundary, set by max depth of Lake Baikal, far exceeds the upstream sediment supply. "Forced" by this downstream boundary condition, the Selenga system should, in theory, be at alluvial grade. Under such a situation, avulsions cease and distributary channels are immobile. On the contrary, field observations are not consistent with this prediction. Lobe avulsions frequently occur at the Selenga Delta, motivating this study of how changing downstream boundary conditions influence alluvial grade.

3. Methods

3.1. Remote-Sensing Analysis

Basin and delta-lobe characteristics of the Selenga River delta, including shoreline position and avulsion locations, were measured using remote-sensing methods. Bathymetry of Lake Baikal and embayments adjacent to the Selenga Delta (Proval and Cherkalovo Bays) were used to measure basin depth and slope (Figures 1b–1db; De Batist et al., 2006; Pavlov et al., 2019; Vologina et al., 2007, 2010). Digital elevation models (DEMs), created by the NASA Shuttle Radar Topography Mission (SRTM), were used to measure topset slope. Manually georeferenced historical survey maps (Figure 2b; n = 4, collected in 1862, 1908, 1956, and 1962; Galazy, 1993; Il'icheva, 2008; Il'icheva et al., 2015; Shchetnikov et al., 2012; Vologina et al., 2007, 2010) and 141 cloud-free Landsat (3, 5, 8) sensor measurements from 1975 to 2019 were used to constrain changes in shoreline and locations of channel avulsion (Figure 2c).

A DEM combining bathymetric and topographic data was created and used to generate elevation profiles that were measured radially based on a semicircular sampling grid with a 180° opening angle, extending from the delta apex to the lake bottom. The datum of the bathymetric and topographic data were relative to the Pulkovo 1942 system and mean global sea level, respectively, and were projected to UTM zone 48°N (De Batist et al., 2006). By setting a 1° grid spacing, a total of 180 radial sampling transects were established (Figure 2c). The grid center was set at the delta apex, defined as the intersection between the axial flow direction of the Selenga River and the adjacent Lake Baikal shoreline (Figure 2c).





Figure 2.



Table 1

Measured Characteristics of the Selenga River Delta and Its Three Lobes

	Western lobe	Central lobe	Eastern lobe	Entire delta
Transect no.	1–65	66–137	138–180	1–180
Receiving basin variables				
Basin slope (\bar{S}_{basin})	0.58–0.63ª	$2.20 \pm 0.60 \times 10^{-2}$	0.58–0.63ª	$1.27 \pm 0.99 \times 10^{-2}$
Basin depth (\bar{H}_{basin}) (m)	1.5 ± 0.4	216 ± 105	2.7 ± 1.0	133 ± 110
Delta lobe variables				
Opening angle $(\bar{\lambda})$	65°	72°	43°	180°
Topset slope (\bar{S}_{fan})	$3.42 \pm 0.36 \times 10^{-4}$	$3.80 \pm 0.42 \times 10^{-4}$	$2.70 \pm 0.42 \times 10^{-4}$	$3.41 \pm 0.58 \times 10^{-4}$
Progradation rate ^b (\bar{R}_{pro}) (m/yr)	12 ± 3	-14 ± 5	19 ± 4	5 ± 4
Lobe radius (\bar{R}) (km)	17.6 ± 0.6	16.7 ± 0.6	19.9 ± 0.9	17.9 ± 1.5
Initial lobe radius (\bar{R}_0) (km)	15.0 ± 0.4	17.7 ± 1.6	15.6 ± 2.3	16.0 ± 2.2
Change in lobe radius $(\Delta \bar{R})$ (km)	2.7 ± 0.7	-1.0 ± 1.3	3.8 ± 2.9	1.4 ± 2.9
Fraction of flux (\bar{F})	$43.6\% \pm 10.9\%$	$16.1\% \pm 4.9\%$	$40.3\% \pm 8.6\%$	100%
Number of outlets (N)	15	9	8	32
Number of avulsion nodes (N)	6	5	3	14
Distributary channel variables				
Water surface slope (\bar{S}_{ws})	$2.24 \pm 0.04 \times 10^{-4}$	$1.74 \pm 0.11 \times 10^{-4}$	$1.84 \pm 0.03 \times 10^{-4b}$	$1.97 \pm 0.26 \times 10^{-4}$
Channel bed slope (\bar{S}_b)	$1.88 \pm 0.41 \times 10^{-4}$	$1.05 \pm 0.33 \times 10^{-4}$	$1.65 \pm 0.51 \times 10^{-4b}$	$1.49 \pm 0.52 \times 10^{-4}$
Bankfull depth $(\tilde{H}_{bf})^{c}$ (m)	$2.7 \pm ^{1.3}_{0.2}$	$2.5 \pm 0.3_{0.7}$	$2.3\pm_{0.4}^{0.4}$	$2.5\pm_{0.4}^{0.6}$
Bankfull width $(\tilde{B}_{bf})^{c}(m)$	$141 \pm \frac{45}{35}$	$45\pm^{20}_{12}$	$122 \pm \frac{28}{21}$	$106\pm^{44}_{24}$
Channel length $(\tilde{L}_c)^c$ (m)	$1,600\pm_{680}^{1,190}$	$1,650\pm^{2,770}_{990}$	$1,480\pm_{810}^{2,220}$	$1,570\pm^{1,620}_{790}$

Note. Values in this table are mean \pm one standard deviation (σ).

^aAngle of repose at 30°-32° (Piliouras et al., 2017; Wang et al., 2019). ^bMean ± 95% confident interval. ^cMedian ± 75th and 25th percentiles.

3.1.1. Measuring Basin and Delta Characteristics: Slope and Depth

Basin slope (S_{basin}) was measured between the delta shoreline and location of maximum curvature of the bathymetric profile (Figure 3). Basin depth (H_{basin}) was defined as the water depth at the location of maximum curvature. For earthquake-impacted (subsided) regions of the delta, basin depth was defined as water depth of the adjacent embayments (Figure 3). To measure solely land elevations, channel pixels (mapped during moderate water discharge, $Q_w = 1$, 100 m³/s) are excluded from SRTM data. Topset slope (S_{fan}) was measured from the delta apex to the shoreline along sampling transects.

3.1.2. Quantifying Shoreline Change

Historical maps and satellite images were used to document the shoreline position of the delta. Shorelines were traced manually from georeferenced historical maps in ArcGIS. For Landsat images, land and water were differentiated using a modified Normalized Difference Water Index (MNDWI), by combining shortwave near-infrared and green bands (Xu, 2006). Shorelines were then extracted automatically from the MNDWI images and manually checked for quality (Moodie et al., 2019). Delta radius was measured as the distance between shoreline and apex for the 180 transects per Landsat image. Annual mean delta radius (\bar{R}) was used to calculate long-term mean progradation rates (\bar{R}_{pro}) over the period of 1862–2019 via a linear relationship between time and shoreline

Figure 2. (a) Map of active faults near the Selenga Delta, produced based on data from Lunina et al. (2014). (b) Digital elevation model of the landscape produced from the NASA Shuttle Radar Topography Mission data. (c) Map of shoreline positions from 1862 to 2019. Deltaic shorelines are extracted from images collected by Landsat missions (3, 5, 8) and historical surveys, spanning 157 years, from 1862 to 2019. A semicircular sampling grid, centered at the delta apex (white diamond), is used to measure attributes of the delta and basin morphology. A total of 180 radial sampling transects, spaced at 1° lobe opening angle (Θ) and originated from the delta apex, are used to make profiles in Figures 7 and 8 (solid red line as example). $\Theta = 0^{\circ}$ at the westernmost transect and $\Theta = 180^{\circ}$ at the easternmost transect. Dashed dark blue line represents the 1862 shoreline.





Figure 3. Sketch of an idealized delta used to calculate lobe volume (after Muto et al., 2016). A range of topset slopes (S_{fan}) and sediment thicknesses (*h*) at the delta apex were used to calculate sediment volume since 1862 (see main text). Note that the topset slope in 1862 $(S_{fan,f0})$ is greater than the topset slope at present (S_{fan}) , with respective sediment thicknesses.

positions (Moodie et al., 2019). Similarly, decadally averaged position were calculated ($\bar{R}_{pro.d}$). Note that data availability is sparse during the period of 1862–1986 (i.e., prior to the Landsat 5 mission). As a result, two measurements of mean radius during this period were spaced by 90 and 20 years, respectively. For the period of 1986–2019 (Landsat Missions 5 and 8), measurements of decadal mean radius were calculated at the correct time interval (i.e., 10 years). Finally, total change in delta radius ($\Delta \bar{R}$) was calculated by differencing shoreline positions for 1862 and 2019.

3.1.3. Identifying Avulsion Sites

To identify avulsion locations, 141 MNDWI images were stacked to generate a water occupation frequency map, an index defined as the fraction of time that a given spatial location (image pixel) is occupied by water (Aminjafari et al., 2021; W. Kim et al., 2006; Piliouras et al., 2017; Straub et al., 2013). This index was then normalized by its maximum value, yielding a normalized water occupation frequency map (NWOF). In particular, new flow pathways had low NWOF values. The NWOF map and Landsat images were examined visually to identify avulsion sites, defined as the formation of a new channel pathway (D. A. Edmonds et al., 2011). Specifically, newly avulsed channels must have a direct connection to a basin, which may include the shoreline of the delta (Lake Baikal), or wetland lakes within the delta plain. Local channel avulsions that rejoin downstream channel are not included for this analysis.

3.2. Field Measurements

Width and depth were measured in seven major distributary channels of the Selenga Delta, using a LOWRANCE single-beam sonar to collect cross-sections over low to bankfull flow conditions during three field expeditions from 2014 to 2018 (60 transects total; Figure 1b; Dong et al., 2016, 2019, 2020). At each location, water surface, channel bank, and

bed elevation were measured using a JAVAD differential Global Navigation Satellite System. These transects were spaced 2.5–4 km apart (Figure 1b). In 2018, water and sediment discharge at 16 sites, located in the same position as the previous surveys, were monitored for 2.5 months to measure flow partitioning in the delta distributary network ($Q_w = 900-2,300 \text{ m}^3$ /s; Figure 1b; Dong et al., 2020).

3.3. Distinguishing Delta Lobes

A graph theory approach is used to identify delta lobes (Dong et al., 2020). Steady-state flux of the Selenga Delta channel network is approximated using a rooted directed acyclic graph (*G*), such that G = (V, E) (Dong & Goudge, 2022; Dong et al., 2020; Tejedor et al., 2015a, 2015b). *V* and *E* are a collection of vertices and links, respectively. Channels are defined as links. Bifurcation and confluence nodes, and channel outlets at the shore-line, are represented by vertices. Link directions correspond to channel flow direction, from the delta apex to the shoreline. Each link contains hydraulic information, such as channel width, and is used to predict flow partition-ing (*F*) for the entire network. A contributing subnetwork is identified for each channel outlet, which contains all the links and vertices that contribute flux to it. Subnetworks can be grouped together as a delta lobe based on the proportion of shared flux using dynamic pairwise dependence (DPD; Tejedor et al., 2015b):

$$DPD_{ij} = \frac{\sum_{u \in S_{ij}} F(u)}{\sum_{v \in S_i} F(v)};$$
(4)

here, S_i is the set of links that belong to subnetwork *i* with vertices of *u*. S_{ij} is the set of links that belong to both subnetwork *i* and *j*, with vertices of *v*. High DPD values indicate that two subnetworks share a large amount of flux. Using this metric, channel outlets and their associated upstream links and vertices are grouped together based on the proportion of shared flux. To account for uncertainties in delta network mapping caused by image



resolution and/or water level variability, the graph-theory-based lobe delineation results are compared to previous geomorphic assessments (Figure 6a; Il'icheva, 2008; Il'icheva et al., 2015), as well as to spatial trends in shoreline progradation rates determined from this study (Section 3.1.2). The final lobe delineation is an average of the three methods.

3.4. Constraining Lobe Volumes

A geometrical framework is used to evaluate change in sediment volume of the delta lobes, following Muto et al. (2016). Assuming sediment balance (V_t) :

$$(1 - \lambda_p) \int_0^t Q_s dt = V_{\rm ac} + V_{\rm aq} = V_t.$$
(5)

 Q_s is the long-term mean sediment discharge in unit of m³/yr, V_{ae} and V_{aq} are the subaerial and subaqueous sediment volumes, respectively. λ_p is the porosity of unconsolidated mixed sand and gravel, $\lambda_p = 0.25$ (Dong et al., 2016; Leopold et al., 1964). Assuming a horizontal basement and a constant sediment discharge, V_{ae} is calculated as a half-cone (Figure 3; Muto et al., 2016; Reitz & Jerolmack, 2012):

$$V_{\rm ac} = \frac{\lambda}{6} h \bar{R}^2, \tag{6}$$

where λ is the delta lobe spreading angle in radians, h is sediment thickness at the delta apex above a datum, and is set as the mean lake level (455 m), $h = \bar{S}_{fan}\bar{R}$, where \bar{R} is the mean delta lobe radius and \bar{S}_{fan} is the mean topset slope. V_{aq} is constrained by a truncated half-cone (Figure 3; Wang et al., 2019):

$$V_{\rm aq} = \frac{\lambda}{2}\bar{H}_{\rm bay}\bar{R}^2 + \frac{\lambda}{2S_{\rm fore}}\bar{H}_{\rm bay}^2\bar{R} + \frac{\lambda}{6S_{\rm fore}^2}\bar{H}_{\rm bay}^3,\tag{7}$$

where \bar{H}_{bay} is the mean water depth in the embayments, and S_{fore} is the foreset slope. For areas impacted by tectonic subsidence, basin slope is equivalent to foreset slope, and assumed to be at angle of repose for fine-grained sediment at $30^{\circ}-32^{\circ}$ (Piliouras et al., 2017; Wang et al., 2019).

Calculating the subaerial sediment volume before the 1862 earthquake (i.e., at such initial time, t_0) requires information about topset slope (S_{fan,t_0}) and sediment thickness (h_{t_0}) at the delta apex, which are difficult values to constrain at t_0 . These variable are also influenced by short-term base level change such as compaction and dam-related lake-level fluctuations. Assuming delta progradation over time, two end-member scenarios bounding possible initial thicknesses and slopes are considered (Figure 3): $h > h_{t_0}$, so that the delta maintains a constant topset slope, $h_{t_0} = S_{\text{fan}} \bar{R}(t_0)$; and $S_{\text{fan}} < S_{\text{fan},t_0}$, whereby sediment thickness at the apex is constant in time, $\bar{S}_{\text{fan},t_0} = h/\bar{R}(t_0)$. To account for variability in base level change, topset slope and sediment thickness from both scenarios are used to calculate change in delta volume and grade index (detailed in Section 3.7). Sediment fill since the 1862 earthquake is calculated for both scenarios as $\Delta V_t = V_{t,2019} - V_{t,1862}$.

3.5. Sediment Discharge

Total sediment load ($Q_{t,pred}$) to the delta is constrained by combing a sediment rating curve and historical hydrograph data, both of which were measured at the main stem from 1938 to 2015 (Figures 4a and 4c; S. R. Chalov et al., 2015; Dong et al., 2020; Pietroń et al., 2018). The long-term mean annual sediment discharge (Q_s) is calculated:

$$Q_s = \frac{1}{t} \int_0^t Q_{t,\text{pred.}} dt, \qquad (8)$$

where t = 78 years is the duration of the historical hydrograph data. Bed material load (Q_{bm}) is calculated by removing the measured mud fraction (grain size <0.0625 mm; silt and clay; 78.7% of the total load) from the measured Q_s , based on the grain size distributions of suspended material measured at the main stem (Figure 4b; S. Chalov et al., 2017; Nittrouer & Viparelli, 2014). In this case, bed-material load (Q_{bm}) includes sand-size sediment that travels as part of both suspension and bed-load transport (Garcia, 2008). Since channel avulsions are driven by bed aggradation, the lower and upper 95% confident intervals of Q_{bm} are used to approximate





Figure 4. (a) Rating curve of total sediment load ($Q_{t,meas.}$) measured for the Selenga Delta main stem (S. R. Chalov et al., 2015; Dong et al., 2020). (b) Grain-size distributions of suspended sediment at the main stem (S. Chalov et al., 2017). (c) Water discharge ($Q_{meas.}$; solid blue line) and predicted total sediment load ($Q_{t,pred.}$; red dashed line) of the Selenga Delta main stem from 1938 to 2015 (Pietroń et al., 2018).

in-channel aggradation rates and to estimate avulsion timescales (Mohrig et al., 2000). Specifically, the fraction of bed material load is assumed to be deposited, thus contributing to deltaic land building.

3.6. Constraining Lobe and Channel Avulsion Timescales

To consider the impact of variable basin depth on delta building processes, the avulsion timescale of the delta lobes $(T_{A,l})$ were calculated as (Muto et al., 2016; Wang et al., 2019):

$$T_{A,l} = \frac{T_{A,l,H_{\text{basin}}\sim 0}}{G_{\text{index}}},$$

$$T_{A,l,H_{\text{basin}}\sim 0} = \frac{\lambda\beta H_{bf,\text{apex}}\bar{R}^2}{2EQ_{\text{brack}}}$$
(9)

where $T_{A,l,H_{\text{basin}}\sim 0}$ is the lobe avulsion timescale at zero basin depth, $H_{\text{bf,apex}}$ is bankfull depth at the delta apex, F is the fraction of sediment discharge that each lobe receives and is constrained using historical and field data (Table 1; S. Chalov et al., 2017; Dong et al., 2020; Il'icheva, 2008), and β is a coefficient that describes the fraction of in-channel aggradation required to setup an avulsion relative to the mean flow depth, and varies between 0.3 and 1 (Chadwick et al., 2019; Ganti et al., 2014, 2016; Jerolmack & Mohrig, 2007; Mohrig et al., 2000; Moodie et al., 2019; Moran et al., 2017). β is unconstrained, so T_A is calculated for a range of values, from 0.3 to 1. The delta-lobe avulsion timescales are derived using a geometric model (Figure 3; Wang et al., 2019), which does not explicitly track locations of avulsed lobes, but instead assumes that the avulsion of the trunk (primary) channel produces a lobe at a different location on the delta plain. This assumption is consistent with field observations at the Selenga Delta (Shchetnikov et al., 2012).



Terraces exist near the delta apex (Dong et al., 2019; Gyninova & Korsunov, 2006). These terraces are thought to have been generated due to relative upward vertical motion along normal faults and/or channel incision following a lobe avulsion in response to tectonic lowering of the delta (Gyninova & Korsunov, 2006; Mats & Perepelova, 2011). Specifically, dropdown of the delta along its shoreline fringes and subsequent channel avulsions drive incision upstream, forming these terraces near the delta apex. Stage and elevation surveys by Dong et al. (2019) revealed that the modern bankfull stage is 0.33 ± 0.19 m below the bank terrace surfaces, consistent with Gyninova and Korsunov (2006), who also documented terraces that are 0.5–2.5 m higher than flood stage. Therefore, H_{bf} is modified by terrace height to account for the distance between channel bed and terrace surface (Equation 9).

For smaller-scale distributary channels downstream of the terraces, the characteristic channel avulsion timescale $(T_{A,c})$ is calculated as (Reitz et al., 2010):

$$T_{A,c} = \frac{\beta \bar{L}_c \bar{B}_{\rm bf} \bar{H}_{\rm bf}}{Q_{\rm bm,c}},\tag{10}$$

where \bar{L}_c , \bar{B}_{bf} , and \bar{H}_{bf} are mean channel length, bankfull width, and depth measured from distributary channels within each delta lobe, respectively. $\bar{Q}_{bm,c}$ is the mean bed material load per channel:

$$\bar{Q}_{\mathrm{bm},c} = \frac{Q_{\mathrm{bm}}F}{N},\tag{11}$$

where N is the number of outlets for each lobe and F is the fraction of water and sediment discharge that each lobe partitions relative to the main river (S. Chalov et al., 2017; Dong et al., 2020).

3.7. Constraining Uncertainties in Delta Lobe and Basin Variables

Predictions of lobe avulsion timescales require knowledge of long-term sediment discharge ($\sim 10^3$ years), which is difficult to constrain, giving that the historical record is only seven decades. Even these data show that sediment discharge has been declining (about 50%) in the Selenga Delta over the last 30 years (S. R. Chalov et al., 2015). It has been proposed that the reasons for this decline in long-term sediment discharge are because of a reduction in erosion rate due to declining agricultural activities and a possible change in regional hydroclimate (S. R. Chalov et al., 2015). Additionally, it is also necessary to constrain original basin water depth at onset of the earthquakes that formed Proval and Cherkalovo Bays. Mud from the Selenga River has been filling these embayments. For example, at Proval Bay, the thickness of post-earthquake sediment fill ranges 0.5–3.6 m (Shchetnikov et al., 2012; Vologina et al., 2010).

To address these problems, a Monte Carlo approach was used to account for stochasticities in delta lobe and basin variables, such as shoreline position, as well as uncertainties in data collection and calculation. Specifically, probability distributions of delta lobe and basin variables were generated (i.e., parameters in Equations 1 and 5–11), and measured from the 180 survey transects (Figure 2c). These variables were randomly sampled 1×10^7 times to generate probability distributions of sediment volume (ΔV_t), grade index (G_{index}), lobe and channel avulsion timescales ($T_{A,l}$ and $T_{A,c}$, respectively) for each delta lobe via Equations 1 and 5–11. The full distribution, as well as the median and 25th and 75th percentiles (quartiles one and three) are reported in discussions below.

4. Results

4.1. Identification of Delta Lobes

A total of 32 vertices are identified as outlets using the graph-theory framework, as they are connected to Lake Baikal or to a surrounding embayment (Figure 5a). Outlets are indexed consecutively and clockwise, starting with the westernmost location (Figure 5a). A subnetwork is identified for each outlet and is compared to its 31 neighbors based on the proportion of shared flux (Figure 5a), yielding a 32×32 DPD matrix. Two distinct populations emerge from the probability distribution of DPD, separated by a threshold value, determined using the otsuthresh tool in MATLAB, DPD = 0.68 (Figure 5c). For the DPD matrix, values are necessarily 1 along the diagonal, as the subnetworks are compared to themselves. Regions of symmetry along the diagonal that contain high DPD values (DPD > 0.68) indicate subnetworks that share more than 68% of influx (Figure 5b).





Dynamic Pairwise Dependence (DPD)

Figure 5. (a) Examples of subnetworks on the Selenga Delta, differentiated using a graph theory framework (Tejedor et al., 2015a). (b) Dynamic pairwise dependence (DPD) matrix used to distinguish lobes within the delta network. Rows and columns are set by the number of delta outlets (subnetworks). DPD values represent the proportion of flux shared between two subnetworks. Regions of symmetry along the diagonal represent a high proportion of shared flux. Interpreted delta lobes are highlighted by boxes with thick outlines. Color scheme of the lobes are consistent for subsequent figures. (c) Probability distribution of DPD values. Two populations emerged, separated by a threshold value, DPD = 0.68.

Using the threshold value of DPD = 0.68, three lobes are interpreted from the DPD matrix (Figure 6). Identified lobes include a western lobe, consisting of outlets 1–18, and an eastern lobe, consisting outlets 24–32; there is no predicted flux shared between the two lobes (Figure 6a). Subnetworks (outlets) 19–23 share flux with the entire delta, and are therefore grouped together and classified as a central lobe. This interpretation of lobes agrees with previous assessments (Figure 6a; Il'icheva, 2008; Il'icheva et al., 2015), as well as with spatial trends in shoreline progradation rates (Figure 6b). Boundaries between the lobes are set at opening angles $\Theta = 65^{\circ}$ and 137°, which are the mean opening angles measured based on the three described methods for distinguishing lobes (Figure 6c).

4.2. Remotely Sensed Data

4.2.1. Basin and Delta Characteristics: Slope and Depth

Bathymetric data analyses indicate that basin slope and depth are highly variable for the three Selenga Delta lobes (Figure 7, Table 1). The central lobe has a basin slope of $2.20 \pm 0.60 \times 10^{-2}$. The western and eastern lobes are surrounded by embayments, and therefore do not have clear division between delta topset and foreset (Figures 7a and 7c). For these two lobes, basin slope (i.e., foreset slope) is assumed to be the angle of repose for fine-grained sediment, $30^{\circ}-32^{\circ}$ (Piliouras et al., 2017; Wang et al., 2019). Basin depth of the central lobe is 216 ± 105 m (Figure 7e). For the western and eastern lobes, embayment bathymetry reveals a mean depth of Cherkalovo Bay at 1.5 ± 0.4 m, and Proval Bay at 2.7 ± 1.0 m, respectively (Figures 7d and 7f).

Analysis of the NASA SRTM data show that topset slopes are variable for the three lobes (Figures 8a–8c, Table 1). The eastern lobe maintains the shallowest topset slope $(2.70 \pm 0.42 \times 10^{-4})$. The topset slope of the central lobe is $3.80 \pm 0.42 \times 10^{-4}$, 41% steeper than the eastern lobe. The topset slope of the western lobe is $3.42 \pm 0.36 \times 10^{-4}$. Based on field surveys of the seven main distributary channels from low to bankfull flow in 2016 and 2018, water surface and bed slopes are largest for channels in the western lobe $(2.24 \pm 0.04 \text{ and } 1.88 \pm 0.41 \times 10^{-4}$, respectively), followed by the eastern $(1.84 \pm 0.03 \text{ and } 1.65 \pm 0.51 \times 10^{-4}$, respectively) and central lobes $(1.74 \pm 0.11 \text{ and } 1.05 \pm 0.33 \times 10^{-4}$, respectively; Table 1, Figure 8f). The central lobe has the steepest topset slope, as well as the largest difference between topset and channel bed slope (Table 1, Figure 8f).

Mean topset elevation profiles are compared between the three lobes (Figure 8d). There is little difference in topset elevation $(\Delta \overline{Z})$ near the apex of the three lobes (Figure 8d). Specifically, values of $\Delta \overline{Z}$ for the central/western lobes, and central/eastern lobes are 0.06 ± 0.38 and 0.29 ± 0.41 m, respectively. However, for regions starting at a distance of 5.0 km downstream of delta apex, the eastern lobe is 1.22 ± 0.53 m higher than the central lobe, thus indicating a lateral gradient, with the central lobe as a relative low. Similarly, for a distance of 10.0 km downstream of the delta apex, the western lobe is 0.42 ± 0.35 m higher than the central lobe. For this study, the area between 5.0 and 10.0 km downstream of apex is termed the region of topset

elevation divergence (Figures 8d and 8e). The mean elevation in this region is 456 m above sea level, and is 1 m higher than mean lake level of 455 m.

4.2.2. Shoreline Change

Analysis of the modern deltaic shoreline position indicates that the eastern lobe has the largest modern radius $(\bar{R} = 19.9 \pm 0.9 \text{ km})$, followed by the western and central lobes ($\bar{R} = 17.6 \pm 0.6 \text{ km}$ and $\bar{R} = 16.7 \pm 0.6 \text{ km}$,





Figure 6. (a) Delta lobes are distinguished using three methods: graph theory (this study), qualitative assessment (II'icheva, 2008), long-term shoreline progradation rates ($\bar{R}_{\rm pro}$; this study). (b) Progradation rates as a function of transect opening angles along the delta. (c) Delta lobe boundaries identified using aforementioned methods. The mean opening angles are $\Theta = 65^{\circ}$ and 137° (that is, solid and dashed lines for the western/central and central/eastern lobe boundaries, respectively).

respectively, Figure 9, Table 1). The long-term mean progradation rate, using shoreline position data from 1862 to 2019, is maximum for the eastern lobe, at 19 ± 4 m/yr. Meanwhile, the progradation rate of the western lobe is 12 ± 3 m/yr, and the central lobe is retreating at 14 ± 5 m/yr (Figures 9a–9c). Decadal mean progradation rate is decreasing for the eastern lobe since the 1862 event, from 23 ± 16 m/yr to -6 ± 10 m/yr (negative rate indicates shoreline retreat; Figure 9d). Similarly, retreat rate of the central lobe decreased from -18 ± 3 to -1 ± 7 m/yr (Figure 9f). During the same time interval, progradation rate of the western lobe increased slightly, from 7 ± 2 to 10 ± 7 m/yr (Figure 9e). Since the 1862 event, the eastern and western lobes have prograded 3.8 ± 2.9 and 2.7 ± 0.7 km basinward, respectively, while the central lobe has retreated 1.0 ± 1.3 km.

4.2.3. Avulsion Sites

A NWOF map shows that the major distributary channels possess high values, indicating water occupation (Figure 10a). Also, this is the case between the distributary channels, where oxbow lakes and abandoned channels are abundant (Figure 10a). Areas of low NWOF values, indicating dry land, are located in the upstream region, near the delta apex, and also adjacent to active channels (e.g., levees; Figure 10a).

A DEM, adjusted to accentuate relatively higher elevation, is compared to a modified map of NWOF showing values <0.05 (indicating less than 5% water occupation frequency; Figure 10b). The comparison shows that regions near the delta apex are both high and dry, due to relict terraces and active levees of the distributary channels (Figure 10b).

Identified channel avulsions are located in areas downstream of the relatively elevated terraced regions. In total, 14 avulsion nodes are identified based on NWOF maps and Landsat images. These nodes are distributed amongst three lobes. Typically, avulsion sites are downstream of the gravel-sand transition, near the region of backwater flow (Dong et al., 2016). Newly avulsed channel pathways usually flow into areas of high NWOF values, indicating avulsions of channels into topographic lows between the major active distributary channels (Figure 10b).

4.3. Field-Measured Distributary-Channel Geometry

Based on field data analysis, channels in the western lobe have the largest median bankfull width and depth $(141\pm^{45}_{35} \text{ and } 2.7\pm^{1.3}_{0.2} \text{ m})$, followed by the eastern and central lobes $(122\pm^{28}_{21} \text{ and } 2.3\pm^{0.4}_{0.4} \text{ m}; 45\pm^{20}_{12} \text{ and } 2.5\pm^{0.3}_{0.7} \text{ m};$ Table 1, Figures 11b and 11c). Coefficient of variations (c_v) for width and depth measurements are largest in the central lobe $(c_v = 0.58 \text{ and } 0.50)$, c_v values are 115% and 39% larger than those of the western and eastern lobes, respectively (Figure 11). In contrast, c_v is smaller in the western and eastern lobes, respectively $(c_v = 0.31 \text{ and } 0.36; c_v = 0.27 \text{ and } 0.39)$.

4.4. Delta Lobe Volumes, Sediment Discharge, and Avulsion Timescales

The calculated volume of sediment deposition above mean lake level since the 1862 earthquake event is highest in the eastern lobe $(0.19\pm_{0.11}^{0.12} \text{ km}^3)$, followed by the western lobe $(0.17\pm_{0.12}^{0.14} \text{ km}^3)$; Equations 6 and 7; Figure 12a). However, since 1862, sediment volume in the central lobe, sequestered below mean lake level, decreased by $0.06\pm_{0.18}^{0.17} \text{ km}^3$ (Table 2, Figure 12a). Mean annual sediment discharge (Q_s) entering the delta at the apex is calculated at $1.10 \times 10^6 \pm 1.06 \times 10^5 \text{ m}^3/\text{yr}$ (Equation 8). This calculated Q_s is consistent with field measurement of total sediment load, $1.47 \times 10^6 \text{ m}^3/\text{yr}$ (S. Chalov et al., 2017). Of this total discharge, mean annual bed material load (Q_{bm}) is $2.35 \times 10^5 \pm 2.26 \times 10^4 \text{ m}^3/\text{yr}$. This value is used to calculate both channel and lobe avulsion timescales ($D \ge 0.0625 \text{ mm}$; 21.3% of the total load; Table 2).



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Figure 7. Water depth profiles from the (a) western, (b) central, and (c) eastern lobes of the Selenga River delta, as measured from the sampling transects. (d and e) Probability distributions of basin water depth measured for each lobe.

Grade index (G_{index}) are variable for the three lobes (Equation 1): $0.67\pm_{0.04}^{0.03}$ for the western lobe, $0.009\pm_{0.003}^{0.007}$ for the central lobe, and $0.050\pm_{0.07}^{0.11}$ for the eastern lobe (Table 2, Figure 12b). The characteristic autogenic lobe avulsion timescales ($T_{A,l}$; Equation 9) are $8,100\pm_{2,300}^{2,800}$, $1.20\times10^{6}\pm_{5.70\times10^{5}}^{8.10\times10^{5}}$, and $9,600\pm_{2,800}^{3,500}$ years for the western, central, and eastern lobes, respectively (Figure 12c). The characteristic channel avulsion timescale ($T_{A,c}$; Equation 10) is $60\pm_{30}^{50}$ years for the western lobe, $20\pm_{10}^{30}$ years for the central lobe, and $20\pm_{10}^{20}$ years for the eastern lobe, which are all significantly shorter than the lobe avulsion timescales (Table 2, Figure 12d). The characteristic lobe and channel avulsion timescales for the entire delta are $T_{A,l} = 12,300\pm_{4,700}^{650,000}$ years and $T_{A,c} = 30\pm_{20}^{60}$ years, respectively (Table 2).

5. Discussions

5.1. Impacts of Tectonic Subsidence on Basin Depth and Delta Avulsion Processes

Tectonic activity around the Selenga Delta generates discrete subsidence events that create shallow embayments along the delta front (Figure 7). As a result, receiving basin depth is variable for each of the three Selenga Delta lobes, affecting avulsion processes operating over temporal scales of multiple centuries to millennia $(>10^2-10^3 \text{ years}; \text{ Figure 13a})$. Avulsions at the delta lobe scale arise predominately due to tectonic subsidence, an allogenic process, which operates at a characteristic length of ~20 km (Table 2, Figure 13a). The 1862 event triggered an avulsion, steering distributary channels into the newly formed Proval Bay, that is, from central to eastern lobes (Figure 2a; Shchetnikov et al., 2012; Vologina et al., 2010). A subsidence event of similar magnitude is suspected to have formed Cherkalovo Bay, driving reorganization of the distributary channels, and diverting water and sediment from the central to western lobes (Shchetnikov et al., 2012).

During the intervening period, distributary channel avulsions occur over a characteristic length scale of ~2 km (i.e., six main channel widths), and over timescale of decades to centuries (Table 2, Figure 13a). These avulsions are situated in the backwater transitional reach, downstream of the gravel-sand transition and alluvial terraces, and thus likely arise due to autogenic processes, including in-channel sediment aggradation caused by lowering downstream shear stress and sediment-transport capacity (Figure 10a; Dong et al., 2016; Nittrouer et al., 2012). Specifically, the median avulsion length scale is $1.40\pm_{0.7}^{0.3}$ of the backwater length scale (Brooke et al., 2022; Dong et al., 2016).

Distributary channels avulse into adjacent low regions between the major active channels. Similar behaviors of compensational filling are also observed for experimental deltas (Figure 10b; Jerolmack & Paola, 2007; Straub



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Figure 8. Topset elevations of the (a) western, (b) central, and (c) eastern lobes of the Selenga River delta measured from NASA Shuttle Radar Topography Mission data for each of the 180 sampling transects. (d) Difference in mean topset elevation $(\Delta \bar{Z})$ between the western/central lobes, and eastern/central lobes, as calculated by subtracting mean profiles. (e) Mean topset elevation profiles for the three delta lobes. (f) Channel bed and topset slopes for the seven distributary channels in the delta (Figure 1b), categorized by lobes.

et al., 2009). Taking the recent Kazanova channel avulsion (1989) as an example, water and sediment discharge are diverted from the eastern lobe into the central lobe, due to the lateral gradient advantage (Figures 2b, 8d and 8e; Aminjafari et al., 2021; Dong et al., 2020). As a result, shoreline progradation rates in the eastern lobe have reduced in time, from 23 ± 16 to -6 ± 10 m/yr (negative value indicates shoreline retreat), while the central lobe has changed from -18 ± 3 m to -1 ± 7 m/yr, indicating that sediment is nourishing the central lobe and limiting shoreline retreat (Table 1, Figures 9e and 9f). In this case, channel avulsions act as a smoothing mechanism to reduce lobe avulsion-generated variability in topography and shoreline roughness (Ganti et al., 2014; Straub et al., 2009).

The scale separation in avulsion lengths has been postulated to be associated with formation mechanism of the distributary channels (Colombera & Mountney, 2022; Jerolmack & Swenson, 2007; Salter et al., 2018; Shaw et al., 2018). Backwater-effect induced distributary channels have length scale of $\sim 10-100$ main channel widths, whereas mouth-bar-induced distributary channels have length scale of $\sim 1-10$ main channel widths (D. Edmonds & Slingerland, 2007; Jerolmack & Swenson, 2007; Shaw et al., 2018). For the Selenga Delta, the separation

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Figure 9. Annual and decadal mean delta radius and progradation rates over time for the (a, d) western, (b, e) central, and (c, f) eastern lobes of the Selenga Delta, since the 1862 earthquake.



Figure 10. (a) Normalized water occupation frequency (NWOF) map, calculated by stacking Normalized Difference Water Index images from 1986 to 2019. Value of 1 (dark blue) indicates areas of continuous water occupation and a value of 0 (light blue) indicates areas of no water occupation. In addition, locations of backwater influence on flow and downstream limits of gravel for the seven distributary channels are shown (Dong et al., 2016). (b) Map showing NWOF values less than 0.05 (indicate dry), overlaid with elevation 1 m greater than mean lake level. The dashed line marks the onset of elevation divergence between eastern/central and western/ central lobes, as shown in Figure 8. Avulsion nodes, original, and new channel pathways are overlain in both panels. White diamond indicate the avulsion node of Kazanova channel (1989).





Figure 11. Measured bankfull (a) depth (H_{bf}) and (b) width (B_{bf}) in channels of the three lobes. Median values \pm quantiles one and three, and coefficient of variance (c_v) are also indicated.

in avulsion length scale is caused by the differences between frequency and magnitude of the allogenic and autogenic avulsion processes. However, regardless of the types of avulsion processes, a majority of the distributary channel bed profiles are continuously adjusting, thus affecting the condition of alluvial grade for the Selenga Delta.



Figure 12. Calculated probability distributions for the three delta lobes: (a) change in sediment volume since the 1862 earthquake (ΔV_i), (b) grade index (G_{index}), characteristic (c) lobe (T_{A_i}), and (d) channel (T_{A_i}) avulsion timescale. Solid lines indicate the median values.



Table 2

Calculated Properties of the Selenga River Delta and Its Three Lobes

	Western lobe	Central lobe	Eastern lobe	Entire delta
Transect no.	1–65	66–137	138–180	1–180
Receiving basin variables				
Tectonic timescale (T_t) (yr)	-	-	-	340-2,600
Delta lobe variables				
Sediment volume $\left(\Delta \tilde{V}\right)$ (km ³)	$0.17 \pm _{0.12}^{0.14}$	$-0.06\pm_{0.18}^{0.17}$	$0.19 \pm 0.12_{0.11}^{0.12}$	$0.12 \pm 0.14_{0.15}$
Total sediment discharge (Q_s) (m ³ /yr) ^a	$4.82 \times 10^5 \pm 4.61 \times 10^4$	$1.77 \times 10^5 \pm 1.70 \times 10^4$	$4.46 \times 10^5 \pm 4.27 \times 10^4$	$1.10 \times 10^6 \pm 1.06 \times 10^5$
Bed material discharge $(Q_{\rm bm})$ (m ³ /yr) ^a	$1.03 \times 10^5 \pm 9.84 \times 10^3$	$3.78 \times 10^4 \pm 3.62 \times 10^3$	$9.50 \times 10^4 \pm 9.10 \times 10^3$	$2.35 \times 10^5 \pm 2.26 \times 10^4$
Alluvial grade $\left(ilde{G}_{ ext{index}} ight)$	$0.67 \pm 0.03 \\ 0.04$	$0.009 \pm _{0.003}^{0.007}$	$0.50\pm^{0.11}_{0.07}$	$0.49 \pm 0.16_{0.48}$
Lobe avulsion timescale $(\tilde{T}_{A,l})$ (yr)	$8,100\pm_{2,300}^{2,800}$	$1.20 \times 10^{6} \pm \frac{8.10 \times 10^{5}}{5.70 \times 10^{5}}$	$9,600 \pm ^{3,500}_{2,800}$	$12,300 \pm \frac{650,000}{4,700}$
Distributary channel variables		5.70×10		
Bed material discharge per channel $(\bar{Q}_{bm,c})$ (m ³ /yr) ^a	$6.85 \times 10^3 \pm 6.56 \times 10^2$	$4.20 \times 10^3 \pm 4.02 \times 10^2$	$1.19 \times 10^4 \pm 1.14 \times 10^3$	$7.64 \times 10^3 \pm 3.77 \times 10^3$
Channel avulsion timescale $(\tilde{T}_{A,c})$ (yr)	$60\pm^{50}_{30}$	$20\pm^{30}_{10}$	$20\pm^{20}_{10}$	$30\pm_{20}^{60}$

^aRating curve predicated values with ±95% confident interval other values in this table are median with ±75th and 25th percentiles.



Figure 13. (a) Composite probability distributions of channel and lobe avulsion timescales for the three delta lobes $(T_{A,c}$ and $T_{A,l}$, respectively), overlain with the range of observed and inferred tectonic timescales $(T_{t} \text{ and } T_{t,i}, \text{ respectively})$. Solid lines indicate the median values. (b) Preserved extremality index (Ω) for the two avulsion processes that operate on the Selenga Delta: channel and lobe avulsions (Ganti et al., 2020). $\Omega \rightarrow 1$ indicates that the sedimentary system preferentially preserves the largest topographic relief (e.g., delta channel at the main stem), while $\Omega \rightarrow 0$ indicates preferential preservation of the most common topographic relief (e.g., distal distributary channels). Error bars indicate standard deviation.



5.2. Impacts of Tectonic Subsidence on Alluvial Grade

Previous experimental studies suggest that a modern river at alluvial grade is most likely to be found in front of a very deep basin (Muto et al., 2016). Due to tectonic subsidence, receiving basin depth is variable around the Selenga Delta, resulting in a range of alluvial grade conditions. The western and eastern lobes are not at alluvial grade, as indicated by the calculated Grade Index, because in-channel sediment aggradation causes distributary channel avulsions (Table 2, Figure 12b). These avulsions occur frequently due to low ratio of accommodation (i.e., shallow embayments) to sediment discharge at the delta front, as supported by a low filling index of $B = \sim 0.03$, calculated using mean subsidence rate between earthquakes of 0.02–0.384 mm/yr (Equation 2; Colman et al., 2003; Liang et al., 2016; Urabe et al., 2004). Geometry and bed profiles of the newly avulsed channels are continuously adjusting. As a result, the difference in western and eastern lobe slopes is small for both the topset and channel bed (Table 1, Figures 8d–8f), while variability in bankfull channel depth and width are also limited (Table 1, Figure 11). Similar patterns of slopes and channel geometry have been observed in experimental deltas that are not at alluvial grade (B. Carlson et al., 2018; Muto et al., 2016). In contrast to the western and eastern lobes, the central lobe is close to alluvial grade ($G_{index} = 0.009 \pm \frac{0.007}{0.003}$, Table 1, Figure 12b). This near-grade condition implies that distributary channels in the central lobe are less likely to avulse. As a result, channel levees have time to develop fully, so that the central lobe possesses a large difference between topset and channel bed slopes, indicating that the main distributary channels have aggraded the topset profile (Table 1, Figures 8d-8f; B. Carlson et al., 2018). The central lobe is also topographically lower than the other two lobes because it receives less sediment since the 1862 earthquake (Table 1, Figures 8d, 8e and 10b). Hydraulic geometry of distributary channels in the central lobe have adjusted to a reduced flow, as it is evident by the fact that they maintain the smallest mean bankfull width and depth of the delta (Table 1, Figure 11).

Findings from this study suggest that a range of channel profiles (i.e., alluvial grade conditions) co-exist on deltas along active margins, implying a range of sediment-transport states to the channel mouths. For example, channels at alluvial grade could be in a state of bypass, whereby sediment is delivered to the foreset, and channels that are not at alluvial grade could rework relict deltaic deposits via avulsion and migration, thus building and preserving stratigraphic patterns that are potentially identifiable in the sedimentary record.

5.3. Impacts of Tectonic Subsidence on the Development of Deltaic Stratigraphy

Discrete tectonic subsidence events are expected to affect the development of stratigraphy for the Selenga Delta. We hypothesize that strata from the Selenga system are built by stratal packages that represent the localized downwarped volume produced by the seismic events. Furthermore, discrete stratal packages should be separated by laterally continuous fine-grained sediment deposited within the subsided embayments. Subsequent delta progradations then builds coarse-grained topset and foreset deposits (i.e., clinoforms) over this fine-grained layer. The stacking pattern of such discrete stratal packages are analogous to parasequences, but have a different formation mechanism (Neal et al., 2016). Specifically, whereas parasequences are often interpreted to be driven by eustatic sea-level cycles, stratal packages for the Selenga Delta are caused by tectonic subsidence. This hypothesis is supported by seismic data collected by Colman et al. (2003), showing multiple prograding clinoform units that contain well-defined sigmoidal internal reflections, bounded by uniform thickness reflections, that is, a fine-grained draped unit. These units are interpreted as deposits of delta topsets and are measured in current water depth of 100-400 m (Colman et al., 2003; C. A. Scholz & Hutchinson, 2000). Assuming mean subsidence of 3-4 m per event and 25% porosity of unconsolidated mixed sand and gravel for compaction (Leopold et al., 1964), the depth of these delta deposits could imply 20-100 subsidence events (Lunina & Denisenko, 2020; Shchetnikov et al., 2012; Vologina et al., 2010). The age at the base of the draped unit that overlay these delta deposits is 650 Kyr, thus providing a characteristic recurrence interval of tectonic subsidence at 6,500–33,000 years (Colman et al., 2003; C. A. Scholz & Hutchinson, 2000). While this inferred tectonic timescale (T_t) is longer than the observed tectonic timescale $(T_t = 340-2,600 \text{ years})$, it is comparable to the autogenic lobe avulsion timescale ($T_{A,l} = 12,300 \pm \frac{650,000}{4,700}$ years), supporting the notion that tectonic subsidence predominately controls delta lobe building for the Selenga system (Figure 13a). However, both tectonic subsidence and autogenic processes could trigger lobe avulsions independently, thus both influencing the delta architecture. For example, the trunk (primary) channel could avulse due to in-channel sedimentation (i.e., an autogenic process) prior to a tectonic subsidence event during a prolonged intervening period between earthquakes. In addition, although not very probable, the locations of a tectonic subsidence event and potential avulsion path could coincide spatially. Furthermore, there is an overlap between the inferred tectonic and calculated autogenic lobe avulsion timescales due to the uncertainties and stochasticities in delta lobe and basin variables (Figure 13a).

While continuous subsidence remains an important mechanism for creating accommodation in the receiving basin, this study suggests that discrete subsidence events can complicate the interpretations of deltaic stratigraphy in tectonically active areas. Specifically, average vertical displacement of the discrete subsidence events is greater than the average distributary channel depth of the Selenga Delta (3 m). As a result, the downdropped delta deposit is not accessible to fluvial reworking and could be completely persevered in the stratigraphic record. In contrast, continuous subsidence rate at the Selenga Delta is 0.02–0.384 mm/yr (Urabe et al., 2004), which makes the associated strata prone to morphodynamic reworking. It may still be challenging to distinguish the signals of discrete and continuous subsidence in the actual stratigraphy, because the downdropped strata could contain previously reworked delta deposits. However, with detailed subsurface data, it is possible to use the fault surface as a bounding surface to differentiate the two subsidence styles.

Similar style of subsidence and preservation like the Selenga Delta is observed in other active rift basins, such as Lake Malawi and Tanganyika near the East African Rift (C. Scholz et al., 1998). Conventional assumption of time-continuous subsidence in analyzing deltaic stratigraphy would lead to interpretation of the stacked delta topset deposits observed in these systems as a results of lake level fall (Urabe et al., 2004). However, findings from this study suggest that such stratal patterns could also emerge due to tectonic subsidence. Herein, we suggest that future studies of deltaic stratigraphy along active margins use indicators of discrete subsidence events (earth-quakes), such as soft-sediment deformation structures (Tanner et al., 2011) and signatures that indicate rapid organic carbon burial (Leithold et al., 2016), to guide stratigraphic and tectonic interpretations.

The hierarchical avulsion processes of the Selenga Delta are expected to affect size of the sedimentary structures persevered in stratal packages. We use the preserved extremality index (Ω) to assess the effect of morphodynamic reworking on the characteristic channel dimensions (i.e., sand body sizes) preserved within each package (Ganti et al., 2020), calculated based on the two levels of morphodynamic hierarchy that modify regional relief of the Selenga Delta: distributary channel and delta lobe avulsions, respectively. The calculated preserved extremality indices are $\Omega = 0.54 \pm 0.18$ and 0.64 ± 0.30 , for channel and lobe avulsions, respectively, indicating that the hierarchical processes could preferentially preserve deeper channels within each stratal package (Figure 13b). Hence, preserved channel sand bodies may be very similar in size (3–4 m deep), contrasting with the distribution found for the modern channels, which possess variable width and depth (one order of magnitude difference), ranging from 10–330 and 0.3–7.0 m, respectively (Figure 11; Dong et al., 2016, 2019). The predicted channel patterns occur because the lobe avulsion timescales are much longer than the channel avulsion timescale ($T_{A,I}$, $T_{A,c} = 40\pm_{26}^{91}$), and so distributary channels are able to rework relict deposits during the quiescent period between impactful earthquakes (Table 2; Ganti et al., 2020). However, future work to obtain high-resolution subsurface data is necessary to validate predictions of preservation of sedimentary structures for the Selenga Delta.

6. Conclusions

In this study, field and remotely sensed delta-lobe and receiving basin characteristics from the Selenga Delta, in Lake Baikal, Russia, are used to assess the effects of tectonic subsidence on basin depth and delta lobe building. For the Selenga Delta, discrete tectonic subsidence events modify basin depth around the coastline by downdropping a portion of the topset (30% of the modern subaerial delta area) below mean channel depth (3 m). The recurrence interval of these impactful events is shorter than autogenic lobe avulsion timescales (340–2,600 years vs. 12,300 years, respectively). Thus, lobe avulsion is triggered predominately by tectonic subsidence, an allogenic process, whereby water and sediment flow are attracted to the newly formed accommodation (partially subsided lobe) due to a regional gradient advantage. This finding suggests that tectonic subsidence is a mechanism that prevents the Selenga Delta to reach alluvial grade, reconciling with the theoretical prediction that the delta should have been "forced" into grade by the downstream boundary: ~1,600 m deep Lake Baikal. During quiescent periods between the subsidence events, channel-scale avulsion occurs more frequently (30 years) due to an autogenic process: in-channel sediment aggradation caused by a backwater effect. As a result, water and sediment are dispersed to topographic lows between the active channels and to the shoreline, generating a semicircular delta geometry. Each subsidence event is expected to be preserved as a discrete stratal package that records evidence of morphodynamic reworking by channel avulsion, leading to preferential preservation of deeper channels. As



rift basins are common sediment sinks, results from this study indicate basin modeling in tectonic active regions should consider the effects of discrete subsidence events and spatial heterogeneous receiving basin depth when considering stratigraphic models.

Data Availability Statement

An open access data repository (https://doi.org/10.5281/zenodo.7415691) hosts the following data used in this study: Selenga Delta shoreline coordinates from 1862 to 2019 and a NWOF map; Bathymetric data of Lake Baikal, Proval and Cherkalovo Bays; shapefiles of Selenga Delta channel network and avulsion locations. Sediment discharge data is extracted from S. R. Chalov et al. (2015) (https://doi.org/10.1007/s12665-014-3106-z), Chalov et al. (2017) (https://doi.org/10.1007/s10113-016-0996-1), and Dong et al. (2020) (https://doi.org/10.1029/2020WR027199). Water discharge data is gathered from Pietron et al. (2018) (https://doi.org/10.1002/hyp.11414) and Dong et al. (2020). Locations of faults in the Selenga Delta were acquired from Lunina et al. (2014) (https://doi.org/10.1016/j.gsf.2013.12.006). Hydraulic data is gathered from Dong et al. (2020). SRTM DEM and Landsat data used here are available for download from https://earth-explorer.usgs.gov.

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