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Zaron, Edward, "Topographic and Frictional Controls on Tides in the Sea of Okhotsk" (2017). *Civil and Environmental Engineering Faculty Publications and Presentations*. 437.

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Topographic and Frictional Controls on Tides in the Sea of Okhotsk

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Abstract

The sensitivity of barotropic tides to bottom topography and frictional parameters has been studied in a model for the Sea of Okhotsk. This region was chosen because of the paucity of bathymetry data and the possibility of using satellite altimeter data to better identify the bottom topography using variational inverse methods. The sensitivity was studied using both the direct and adjoint sensitivity. In the former approach, perturbations to the nominal model were applied to examine their impact; in the latter approach, the sensitivities were computed using the adjoint of the tangent linearization of the dynamical model. It is found that small-scale coastal near-resonant amplification controls the tidal dynamics, and the sensitivity of the solutions is dominated by topography in these regions, far exceeding the influence of other factors. Consequently, the tidal dynamics and resonant amplification creates a non-local relationship between water level and bottom topography and leads to a linear dependence of measurements upon a very few degrees of freedom. The results indicate severe limitations on inverse approaches for

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identification of topography, and add to the rationale for the collection and sharing of high quality bathymetry data to enable improved ocean modeling.

Keywords:

tide, adjoint sensitivity, bottom topography

1. Introduction

Numerical modelers of the coastal ocean have long recognized the importance of bottom topography for making accurate models of tides and water levels (Lefevre et al., 2000). The bottom topography is here defined as the two-dimensional field of water depth, i.e., the vertical distance between the mean water surface and the material bottom. The term *bathymetry* is sometimes used to refer to *bottom topography*; however, here the usage of *bathymetry* shall be restricted to refer to measurements of water depth via hydrographic surveys (Organization, 1998), rather than referring to the two-dimensional depth field.

Bottom topography influences the flow fields through several distinct mechanisms. First, bottom topography defines the lower material boundary of the ocean, which provides kinematic constraints on the flow. Second, at a given location, the water depth is related to the mass of the water column and determines the acceleration resulting from a horizontal force; the speed of wave propagation is thus set by water depth. And, finally, the ocean bottom is obviously the locus of boundary layer processes, and these support tangential stresses that may influence entire the water column. Although the detailed representation of these physics differs depending on the setting, their manifestations on wave processes, material transport, vorticity

21 dynamics, and dissipative processes are frequently significant.

22 It can be a challenge for ocean modelers to obtain the topography needed
23 to create credible ocean simulations. The availability of bathymetry obtained
24 via ordinary or multibeam sonar varies greatly as a function of geographic lo-
25 cation (Wessel and Chandler, 2011), and the accuracy of older measurements
26 is not well characterized (Jakobsson et al., 2002; Marks and Smith, 2008).
27 In data from the pre-satellite era, the navigational or geolocation error (20
28 km being a typical magnitude, Smith 1993) contributes the most to uncer-
29 tainty in depth, in direct proportion to the slope of the bottom (Jakobsson
30 et al., 2005). At the interface between the continental shelf and slope, the
31 error in depth can be 100’s of meters (Marks and Smith, 2005; Inazu et al.,
32 2009). Other sources of error include the uncertain vertical datum, when
33 the measured depth is significantly altered by tides or other time-variable
34 water depth which is not corrected, error due to the non-vertical orientation
35 of the sonar beam, and error in the speed of sound used to convert travel
36 time to distance (Marks and Smith, 2008). Smith and Sandwell (1994) pio-
37 neered the use of the marine gravity field for inferring bottom topography in
38 the deep ocean, where sediments are uniform and the relationship between
39 gravity and topography can be inferred. This approach is less accurate over
40 the continental shelves and regions of complex geology (Marks and Smith,
41 2012), where dense in situ data are necessary to control the gravimetry-based
42 estimates (Marks et al., 2010; Marks and Smith, 2012).

43 In order to deal with these uncertainties, it is typical in the course of
44 model development to examine the sensitivity of the solution to changes in the
45 source of the topography and the degree of smoothing, as well as factors such

46 as grid resolution (Hirose et al., 2001). And, sometimes, the topography is
47 calibrated or adjusted to improve the model performance (Ten-Brummelhuis
48 et al., 1993; Lyard and Genco, 1994; Hirose, 2005; Cea and French, 2012).

49 The present study was motivated by the hypothesis that an inverse model-
50 ing approach could usefully improve the accuracy of bottom topography over
51 continental shelves at locations where bathymetry is sparse and gravimetrically-
52 derived topography is inaccurate. The barotropic tides are probably the
53 most-precisely measured and predictable phenomena involving the oceans at
54 continental-shelf scale (Stammer et al., 2014), so the inverse approach was de-
55 veloped by assimilating altimeter-derived water elevation measurements into
56 a dynamical tide model. The technical approach involved re-working the
57 implementation of Zaron et al. (2011) to the frequency domain by augment-
58 ing the approach of Egbert et al. (1994) with topography as a distributed
59 control parameter. Poor results with the inverse approach and unexpected
60 sensitivity to the spatial correlation model for the topographic error led to the
61 more fundamental investigation of model sensitivity described in the present
62 article.

63 This paper is organized as follows. In Section 2, the concepts of the for-
64 ward and adjoint sensitivity are defined and their application to a barotropic
65 shallow water model of tidal dynamics is reviewed. In Section 3, an overview
66 of the factors influencing the accuracy of tide models in the Sea of Okhotsk is
67 conducted in order to define the relative importance of bottom topography,
68 grid resolution, frictional parameterization, and non-linear tidal interactions.
69 In Section 4, the adjoint sensitivity is used to quantify the sensitivity of tidal
70 elevation to topography and friction, and the linear dependence of the ad-

joint sensitivity functions is examined. In Section 5, the findings of Section 4 are related to a previous idealized study that emphasized the significance of tidal resonances near the coastline, and other implications of the results are discussed. Section 6 summarizes the results.

2. Methodology

2.1. Tide Model and Dynamics

The tidal dynamics are expressed in terms of the horizontal volume transport vector, $\mathbf{U}$, comprised of zonal and meridional components (U, V) , and the water surface elevation, η , which are functions of latitude, θ , longitude, ϕ , and time, t . The time-varying fields $(\mathbf{U}, \eta)$ are expanded as the sum of contributions from $k = 1, \dots, N_c$ astronomical tidal constituents in the present case,

$$\mathbf{U}(\phi, \theta, t) = \sum_{k=1}^{N_c} \text{Re}[\mathbf{U}^{(k)}(\phi, \theta) \exp(-i\omega_k t)] \quad (1)$$

$$\eta(\phi, \theta, t) = \sum_{k=1}^{N_c} \text{Re}[\eta^{(k)}(\phi, \theta) \exp(-i\omega_k t)], \quad (2)$$

where $(\mathbf{U}^{(k)}, \eta^{(k)})$ are complex-valued harmonic constants for the transport and surface elevation fields, and ω_k are the frequencies of tides. In the present case, no more than $N_c = 8$ constituents shall be considered, corresponding to the M_2 , K_1 , S_2 , O_1 , N_2 , P_1 , K_2 , and Q_1 tides, for $k = 1, \dots, 8$, respectively. Henceforth the notation indicating a particular frequency (k) shall be omitted except when necessary in expressions involving more than a single frequency.

Tidal dynamics are governed approximately by the Laplace Tidal Equations (LTE) augmented with a linear approximation of the quadratic bottom

92 drag law (Snyder et al., 1979) and a modified astronomical tide generat-
 93 ing force (ATGF) which accounts for solid Earth loading and ocean self-
 94 attraction (Egbert et al., 1994). The LTE are given by,

$$-i\omega \mathbf{U} + f \times \mathbf{U} + gH\nabla(\eta - \Phi) + C_d u_f \frac{\mathbf{U}}{H} = 0 \quad (3)$$

$$-i\omega\eta + \nabla \cdot \mathbf{U} = 0 \quad (4)$$

95 where $f = 2\Omega \sin \theta \hat{k}$ is the local vertical component of the Coriolis parameter,
 96 $g = 9.81\text{m/s}^2$ is gravitational acceleration, H is the bottom topography,
 97 $C_d = 2.5 \times 10^{-3}$ is the bottom drag coefficient, and u_f is an estimate for
 98 the time-average near-bottom current speed. The domain, denoted $\mathcal{D}$, has
 99 boundary, $\partial\mathcal{D}$, comprising closed (material) and open boundary segments,
 100 denoted $\partial\mathcal{D}_1$ and $\partial\mathcal{D}_2$, respectively. Boundary conditions are no-normal flow
 101 on $\partial\mathcal{D}_1$,

$$(U, V) \cdot \mathbf{n} = 0, \quad (5)$$

102 and specified surface elevation, η_d , on $\partial\mathcal{D}_2$,

$$\eta = \eta_d. \quad (6)$$

103 The LTE are thus forced by the modified ATGF, Φ , and by open boundary
 104 conditions on tidal elevation η . In spherical polar coordinates the gradient
 105 and divergence operators are given by,

$$\nabla\eta = \left(\frac{1}{a \cos \theta} \frac{\partial\eta}{\partial\phi}, \frac{\partial\eta}{a\partial\theta} \right), \quad (7)$$

106 and

$$\nabla \cdot \mathbf{U} = \frac{1}{a \cos \theta} \frac{\partial U}{\partial\phi} + \frac{1}{a \cos \theta} \frac{\partial(V \cos \theta)}{\partial\theta}, \quad (8)$$

107 where $a = 6.7308 \times 10^3$ km is the radius of the Earth.

108 In order to model the frictional coupling of different frequencies, u_f is
 109 computed from the tidal currents, $\mathbf{U}^{(k)}/H$, and non-tidal current, u_0 (Snyder
 110 et al., 1979), as

$$u_f = \left(u_0^2 + \gamma_f \frac{1}{2H^2} \sum_k |\mathbf{U}^{(k)}|^2 \right)^{1/2}. \quad (9)$$

111 The optimal value of the coefficient γ_f is related to the frequency con-
 112 tent of the tides and may be found by analysis of the quadratic bottom
 113 stress (Dronkers, 1964; Snyder et al., 1979; Lavelle and Mofjeld, 1983). Here
 114 the values $\gamma_f = 1$ and $u_0 = 0.25$ m/s are used so that u_f is simply the root-
 115 mean-square current speed accounting for the resolved tides and a constant
 116 non-tidal current. Even with this approximation, the solution of (3)-(6) is
 117 iterative, with u_f being evaluated from currents computed at the previous
 118 iteration. In practice, 5 iterations are typically used to produce stable esti-
 119 mates of u_f which differ by less than a few percent from previous values.

120 Note that $(\mathbf{U}, \eta)$ depends on tidal frequency (k), but H and C_d do not
 121 depend on (k). Thus, in addition to the coupling through u_f , the equations
 122 for $(\mathbf{U}^{(k)}, \eta^{(k)})$, (3)-(4), are implicitly coupled via H and C_d . Additional
 123 nonlinearity in the shallow water equations has been omitted from the LTE
 124 without detailed justification. Later, in Section 3, some of these nonlinearities
 125 shall be examined in more detail by using a specific solution of the LTE to
 126 diagnose these terms.

127 2.2. Forward Sensitivity

128 Suppose the topography and/or drag coefficient are perturbed from back-
 129 ground fields, $H = \bar{H} + H'$ and $C_d = \bar{C}_d + C'_d$. Assuming the correspond-
 130 ing background fields, $(\bar{\mathbf{U}}, \bar{\eta}, \bar{H}, \bar{C}_d)$, are solutions of equations (3)-(4), the

131 $(\mathbf{U}', \eta', H', C'_d)$ perturbations approximately satisfy the tangent-linearization
 132 of the LTE,

$$\begin{aligned}
 -i\omega \mathbf{U}' + f \times \mathbf{U}' + g\overline{H}\nabla\eta' + \overline{C}_d u_f \frac{\mathbf{U}'}{\overline{H}} + gH'\nabla(\overline{\eta} - \Phi) + \\
 C'_d u_f \frac{\overline{\mathbf{U}}}{\overline{H}} - \overline{C}_d u_f \frac{\overline{\mathbf{U}}}{\overline{H}^2} H' &= 0 \quad (10) \\
 -i\omega\eta' + \nabla \cdot \mathbf{U}' &= 0, \quad (11)
 \end{aligned}$$

133 with homogeneous boundary conditions on $\partial\mathcal{D}_1$ and $\partial\mathcal{D}_2$.

134 The typical approach to assessing the sensitivity of a solution to per-
 135 turbations of topography or drag coefficient involves differencing solutions
 136 of (3)-(4) obtained with different H or C_d fields. When either H' or C'_d
 137 are given, the so-called forward sensitivity of $y = (\mathbf{U}, \eta)$ may be directly
 138 computed from the Gâteaux derivative (Andrews and Hopper, 2011),

$$\frac{\partial y}{\partial \tau}(\overline{H}, \overline{C}_d; H', C'_d) = \lim_{\tau \rightarrow 0} \frac{y(\overline{H} + \tau H', \overline{C}_d + \tau C'_d) - y(\overline{H}, \overline{C}_d)}{\tau}, \quad (12)$$

139 where the notation makes it clear that the derivative depends on both the
 140 background state and the direction of the perturbation. The code imple-
 141 menting the tangent-linear LTE has been tested to verify its correctness in
 142 the limit $\tau \rightarrow 0$, but the examples shown use finite-amplitude perturbations
 143 with $\tau = 1$. Although there are significant nonlinear components of the sen-
 144 sitivities in the LTE model, they are not pertinent to the overall conclusions
 145 and will not be discussed below.

146 2.3. Adjoint Sensitivity

147 While the forward sensitivity can be computed for a given perturbation,
 148 (H', C'_d) , it is sometimes useful to consider the sensitivity of a particular

149 quantity to any possible perturbation. This kind of object, the derivative
 150 of a scalar functional of the fields with respect to all possible perturbations,
 151 is often called the *adjoint sensitivity* (Errico, 1997). The adjoint sensitiv-
 152 ity provides information about particular changes in the output fields, say,
 153 elevation at a given point, as a function of arbitrary perturbations, which
 154 are not specified a priori. Generalizing the notation introduced above, let
 155 $x = (H, C_d)$. The perturbation outputs, $y' = (U', \eta')$, can be computed from
 156 perturbation inputs, $x' = (H', C'_d)$, as,

$$y' = \int_{\mathcal{D}} \left[\frac{\partial y}{\partial x} \right] x', \quad (13)$$

157 where the expression $\frac{\partial y}{\partial x}$ is the Fréchet derivative (Andrews and Hopper,
 158 2011), an integro-differential operator. It is difficult to visualize $\frac{\partial y}{\partial x}$ directly,
 159 but it is straightforward to visualize the derivative of a scalar functional of
 160 y ,

$$u_i = \int_{\mathcal{D}} \kappa_i y', \quad (14)$$

161 where, for example, the kernel κ_i might represent a delta-function applied to
 162 the η -component of y , so that u_i measures the water elevation at a particular
 163 point, indexed i , in the domain. The perturbed measurement would then be
 164 given by,

$$u_i = \int_{\mathcal{D}} \left[\int_{\mathcal{D}} \kappa_i \frac{\partial y}{\partial x} \right] x', \quad (15)$$

165 where the expression, $\int_{\mathcal{D}} \kappa_i \frac{\partial y}{\partial x}$, is analogous to the i -th row of a finite-dimensional
 166 Jacobian matrix (Bennett, 1992, 2002).

The object $(\lambda, c) = \int_{\mathcal{D}} \kappa_i \frac{\partial y}{\partial x}$ is the adjoint sensitivity. It contains compo-
 nents, (λ, c) , corresponding to the (H', C'_d) components of x' in equation (13).

When equation (13) is written out for the tangent-linear LTE, one may integrate by parts to obtain the following system to compute (λ, c) for a given measurement kernel κ_i ,

$$i\omega\boldsymbol{\mu} - f \times \boldsymbol{\mu} - \nabla\zeta + \overline{C}_d u_f \frac{\boldsymbol{\mu}}{\overline{H}} = 0 \quad (16)$$

$$i\omega\zeta - \nabla \cdot (g\overline{H}\boldsymbol{\mu}) = -\kappa_i \quad (17)$$

$$\lambda = -\sum_k \text{Re} \left[g\boldsymbol{\mu}^{*(k)} \cdot \nabla(\overline{\eta}^{(k)} - \Phi^{(k)}) - \overline{C}_d u_f \frac{\boldsymbol{\mu}^{*(k)} \cdot \overline{\mathbf{U}}^{(k)}}{\overline{H}^2} \right] \quad (18)$$

$$c = -\sum_k \text{Re} \left[u_f \frac{\boldsymbol{\mu}^{*(k)} \cdot \overline{\mathbf{U}}^{(k)}}{\overline{H}} \right]. \quad (19)$$

167 Boundary conditions on ∂D_2 are,

$$\boldsymbol{\mu} \cdot \mathbf{n} = 0, \quad (20)$$

168 and on $\partial \mathcal{D}_1$ are,

$$\zeta = 0. \quad (21)$$

169 Similar to (λ, c) and (H, C_d) , the adjoint fields $(\boldsymbol{\mu}, \zeta)$ correspond to the phys-
 170 ical fields $(\mathbf{U}, \eta)$; however, unlike (H, C_d, λ, c) , the fields $(\mathbf{U}, \eta, \boldsymbol{\mu}, \zeta)$ depend
 171 on tidal frequency (k) . Note that the adjoint fields also depend on κ_i , the
 172 measurement kernel, and $(\overline{\mathbf{U}}, \overline{\eta}, \overline{H}, \overline{C}_d)$, the background fields; the frictional
 173 speed, u_f , is also a function of $(\overline{H}, \overline{\mathbf{U}})$. Unlike the direct sensitivity, the ad-
 174 joint sensitivity does not depend on (H', C'_d) . The computations involved in
 175 the adjoint system, (16)-(21), are implemented by taking the adjoint of the
 176 finite-difference implementation of the tangent-linear system, (10)-(11).

177 In the examples to be shown below, the measurement kernel, κ_i , cor-
 178 responds to the in-phase or quadrature (i.e., real or imaginary component)

179 harmonic constant of water surface elevation of the k -th tidal frequency at
 180 particular latitude and longitude coordinates. One solves equations (16)-
 181 (19) to compute (λ_i, c_i) for the given κ_i . The perturbation water elevation,
 182 $\eta'(\theta_i, \phi_i)$, for a particular H' is then given by the inner product of λ and H' ,

$$\eta' = \int_{\mathcal{D}} \lambda_i H'. \quad (22)$$

183 Instead of specifying H' or C'_d a priori, one specifies the measurement kernel,
 184 κ_i , and computes the fields (λ, c) which η is sensitive to.

185 **3. Tides and Tide Model Sensitivities in the Sea of Okhotsk**

186 Some of the largest tides in the world are found in the Sea of Okhotsk.
 187 For reference, cotidal charts for M_2 and K_1 tides from the data assimilating
 188 TPXO7.2 model (Egbert and Erofeeva, 2002) are shown in Figure 1, and
 189 these agree closely with the earlier estimates of Kowalik and Polyakov (1998).
 190 Semi-diurnal tides are amplified at Uetskaya Guba (Uda Bay), in the west,
 191 where the M_2 amplitude reaches 3.0m. Diurnal tides are amplified in the
 192 Shelikhov Gulf and in Penzhinskaya Guba (Penzhina Bay), in the northeast,
 193 where K_1 amplitudes exceed 3.0m; although, the TPXO7.2 solution does not
 194 capture the highest values. The large tidal elevations are associated with
 195 strong tidal currents, which cause vertical mixing and play a role in the
 196 generation of water masses in the North Pacific (Shcherbina et al., 2004).

197 To understand the topographic and frictional controls on tides, it is useful
 198 to compare the TPXO7.2 model with purely dynamical (not data assimilating)
 199 models based on slightly different topographic grids and different friction
 200 coefficients. Two independent estimates of topography are used. The first is

201 the Digital Bathymetric Database version 3.0 (DBDB2v30) which is based
 202 on a compilation of digitized bathymetric charts and other data blended
 203 and gridded at uniform 2-minute resolution (Dong Shan Ko, personal com-
 204 munication; Ko 2010). The second is the ETOPO1 topography (Amante
 205 and Eakins, 2009) which, in the Sea of Okhotsk, is a hand-edited version of
 206 the topography inferred by combining sounding data with satellite-altimeter
 207 derived marine gravity data (Smith and Sandwell, 1997). The tidal fields
 208 shown in Figure 2 are computed from a numerical LTE solver on an ap-
 209 proximately 3 km horizontal resolution grid using the two dominant tidal
 210 constituents, M_2 and K_1 , with elevation on the open ocean boundary pro-
 211 vided by TPXO7.2. The numerical implementation is described in detail
 212 in Egbert and Erofeeva (2002).

213 The tides obtained with different bottom topographies differ substantially.
 214 For M_2 , the most apparent difference is the larger tide in the northeast in the
 215 Shelikhov Gulf, an enhancement of resonance caused by the different topog-
 216 raphy. One also notes the large-scale rotation of 30° of the phase around the
 217 amphidromes. Both of the tide models in Figure 2 display much larger K_1 am-
 218 plitudes compared to the TPXO7.2 model, which is not well constrained
 219 by data in the Shelikhov Gulf. The differences between the ETOPO1 and
 220 DBDB2v30 solutions extend throughout the central Sea of Okhotsk, with
 221 K_1 tide in the DBDB2v30 model being much larger than in the ETOPO1
 222 model.

223 Several other tide models were developed using the ETOPO1 topography
 224 in order to assess the significance of the topography relative to model reso-
 225 lution, drag coefficient, and number of tidal constituents included. Table 1

226 summarizes the error statistics of the ETOPO1-based model for different ver-
 227 sions using $N_c = 1, 2, 4$, and 8 constituents corresponding to the constituents
 228 listed in Section 2.1. The tabulated root-mean-square vector error (RMSVE)
 229 is equal to one-half the square root of the sum of the squared errors between
 230 the modeled and observed in-phase and quadrature components of the har-
 231 monic constants; the RMSVE is equal to the root-mean-square of the mean
 232 difference in time of the partial tide. The data in this case come from the
 233 merged records of the TOPEX/POSEIDON, Jason-1, and Jason-2 altimeter
 234 missions, which have been harmonically analyzed at each point along the
 235 satellite ground tracks (Ray, 1998; Carrère et al., 2004), and then averaged
 236 over approximately 150 km segments to obtain values at the orbit cross-over
 237 points (Figure 1a). In order to retain only the highest quality data, data are
 238 not used at points with more than 10% missing data owing to seasonal sea
 239 ice or land-contamination of radiometer data used for the wet-troposphere
 240 correction.

241 Table 1 indicates that the RMSVE is relatively insensitive to both model
 242 resolution and N_c . For $N_c \geq 2$, the RMSVE for M_2 ranges from 3.1 cm to
 243 4.1 cm, which is smaller than the RMSVE difference between the nominal
 244 ETOPO1 solution (3.2 cm) and the TPXO7.2 reference solution (1.3 cm).
 245 Similarly, the K_1 range, 4.7 cm to 6.0 cm, is smaller than the difference
 246 between the nominal ETOPO1 solution (4.8 cm) and the TPXO7.2 refer-
 247 ence solution (2.7 cm). Thus, the ETOPO1 model cannot be substantively
 248 improved compared to the TPXO7.2 model by changing either the grid res-
 249 olution (3 km) or the number of tidal constituents ($N_c = 2$).

250 The ETOPO1-based model has also been run using a range of values for

Table 1: Root-mean-square vector error (RMSVE, units of cm) of models versus observed M_2 & K_1 . For comparison, the RMSVE of the TPXO7.2 model is 1.3 cm for M_2 , and 2.7 cm for K_1 .

N_c	M ₂ RMSVE [cm]			K ₁ RMSVE [cm]		
	4.5 km	3 km	2 km	4.5 km	3 km	2 km
1	3.9	3.1	3.5	n.a.	n.a.	n.a.
2	4.1	3.2	3.7	6.0	4.8	5.4
4	4.0	3.1	3.6	6.0	4.7	5.3
8	4.0	3.1	3.6	5.9	4.6	5.3

the bottom drag coefficient, C_d , from $C_d = 0.5 \times 10^{-3}$ to $C_d = 10 \times 10^{-3}$. The RMSVE of the ETOPO1-based model varies from about 3 to 6 cm over this range (Figure 3). The M_2 RMSVE is optimized at approximately $C_d = 2.25 \times 10^{-3}$, while the K_1 RMSVE is optimized at $C_d = 6.0 \times 10^{-3}$. In both cases, though, the RMSVE changes due to C_d are much less than the changes due to topography. The RMSVE obtained using the DBDB2v30 topography (with the same 3 km resolution, $N_c = 2$, and $C_d = 2.5 \times 10^{-3}$ used for ETOPO1) is more than 2 times larger than the RMSVE of the ETOPO1-based model (Figure 3).

The DBDB2 and ETOPO1 topographic grids are both plausible estimates of topography, given the sparsity of soundings in the Sea of Okhotsk, but the ETOPO1 topography definitely contains smaller-scale features than DBDB2v30 (Figure 4). The difference field reveals that there are also larger-scale differences that extend throughout much of the middle of the Sea (Fig-

ure 5a). Of course, the largest differences occur near where topographic gradients are largest, offshore of the Kuril Islands, but, as will be shown quantitatively, below, these differences occur in deeper water and are less significant to tides within the Sea. The large scale differences in the middle- and eastern parts of the main basin are about 150 m, which is more than 10% of the depth in many places.

Figure 5b shows the sparsity of in situ bathymetry available in the region. The points shown in black are the bathymetric control points used in version 9.1 of the Smith & Sandwell bottom topography, which is contemporary with the ETOPO1 topography from 2009. Comparison with independent bathymetry from the region (50°N,148°E)-(55°N,151°E) (cruise MR06-04 obtained from the Japan Agency for Marine-Earth Science and Technology (JAMSTEC)¹) finds standard errors of about 60 m and 80 m for DBDB2 and ETOPO1, respectively. Thus, even though the DBDB2v30 is much smoother than ETOPO1, it appears to be of comparable accuracy. The ETOPO1 grid, being derived largely from altimetry, contains errors related to the uncertain relation between marine gravity and topography in this region (Marks and Smith, 2012), and without additional validation data it is difficult to verify that the small-scale structure is accurate.

In summary, a comparison of tide models indicates that bottom topography has the greatest impact on the tidal fields, considering a plausible range of changes in physical and numerical model parameters.

Nonetheless, there are other factors which cannot easily be examined

¹<http://www.godac.jamstec.go.jp/>

288 within the context of using a single numerical model. For example, in spite
 289 of the examination of various values for the bottom drag coefficient, it is
 290 difficult to justify any particular choice for C_d . The quadratic drag law is a
 291 bulk model for the turbulent stress at the bottom, which depends on shear
 292 and stratification throughout the water column, non-tidal currents, as well
 293 as the physical characteristics of the bottom itself (Grant and Madsen, 1979;
 294 Xing and Davies, 1995; Perlin et al., 2005). The linearization of the quadratic
 295 drag law introduces additional uncertainty in the value of u_f which can also
 296 be lumped into C_d (Snyder et al., 1979). And, in the Sea of Okhotsk, there
 297 is seasonal sea ice, which adds more friction in the mean (Godin, 1986). The
 298 numeric size of the drag term is expected to be essentially negligible in water
 299 deeper than 50 m, so it is only drag near the coast which is likely significant.
 300 Figure 6 illustrates this quantitatively by plotting a friction number, R_f ,
 301 defined by,

$$R_f^2 = \frac{(C_d u_f |\mathbf{U}^{(k)}|/H)^2}{(\omega_k |\mathbf{U}^{(k)}|)^2 + (C_d u_f |\mathbf{U}^{(k)}|/H)^2}, \quad (23)$$

302 which measures the size of the frictional terms in the momentum equa-
 303 tions (3). This quantity is a function of tidal frequency (k) and the M_2 field is
 304 shown in the figure; the corresponding balance for K_1 is similar (not shown).
 305 The frictional term dominates the term balance in shallow water, but R_f is
 306 never larger than 0.5 in water deeper than 100m (Figure 6). Thus, factor of
 307 two or larger mis-specification of the frictional parameters (C_f , u_0 , and γ_f)
 308 ought to influence the term balance only in the shallow regions shown.

309 The final factor to consider is the validity of other approximations im-
 310 plicit in the LTE, namely, the neglect of the advective nonlinearity and the
 311 neglect of the flooding and drying processes at the coastline. The impact of

312 these approximations can be compared with the drag terms by looking at
 313 the relevant non-dimensional parameters. Advection is quantified with the
 314 bulk Froude number, estimated as the ratio of the friction velocity (which
 315 incorporates all the tidal currents, plus the non-tidal current speed, u_0) to
 316 the shallow water phase speed in the absence of rotation,

$$Fr = \frac{u_f}{\sqrt{gH}}. \quad (24)$$

317 The Froude number is the appropriate measure of the advective nonlinearity
 318 when the pressure gradient is among the dominant terms. This field is illus-
 319 trated in Figure 7b, where it is compared with R_f in a close-up of Udkaya
 320 Guba. The Fr field approaches 0.5 in some of the shallowest regions near the
 321 coast, but these are always contained within the region where friction dom-
 322 inates. Likewise, the vertical excursion number, Z , may be defined which
 323 measures the tidal range in comparison with mean water depth,

$$Z^2 = \frac{1}{2} \sum_k |\eta^{(k)}|^2 / H^2. \quad (25)$$

324 Significant values of Z indicate the potential for inundation within a much
 325 smaller region than encompassed by frictional or advective influences; al-
 326 though, the presence of inundation per se might alter dissipation which could
 327 have larger-scale impacts.

328 4. Adjoint Sensitivities in the Sea of Okhotsk

329 The most direct way to assess the sensitivity of tidal fields to topography
 330 or drag is by computing solutions of the adjoint LTE system. Here this is
 331 done by computing the λ and c fields defined in equations (18) and (19). The

Figure 1: TPXO7.2 tide model for the (a) M_2 and (b) K_1 tides. Tidal amplitude is shown with the color scale, which is different for M_2 and K_1 . Greenwich phase lag is shown with solid lines in increments of 30° . Solid dots in (a) are locations of satellite altimeter data used to compute the error statistics summarized in Table 1.

332 measurement kernel, κ_i , is a delta-function corresponding the measurement
 333 of the in-phase M_2 component of $\eta^{(k)}$ at mid-basin, near 55°N - 150°E .

334 In order to display the large numeric range of values of λ , Figure 8 shows
 335 the base-10 logarithm of $|\lambda|/\lambda_{max}$, where λ_{max} is the maximum value of $|\lambda|$.
 336 The most striking feature of the adjoint sensitivity is that it varies by several
 337 orders of magnitude between the regions of tidal amplification and the mid-
 338 basin location. Thus, the η -field at mid-basin is primarily sensitive to the
 339 values of water depth at rather distant nearshore regions.

340 The sensitivity of η to a given perturbation in H is given by the inner
 341 product written in equation (22). To assess the physical significance of the
 342 sensitivity, let's assume H' is proportional to H , so a more relevant measure of
 343 the sensitivity is the product of λ times H . This base-10 logarithm of this field

Figure 2: ETOPO1- (top) and DBDB2v30- (bottom) based tide models. Note that the color scales differ for M_2 and K_1 but they are the same as in Figure 1.

Figure 3: RMSVE as a function of C_d using ETOPO1 topography for M_2 (solid) and K_1 (dashed). Points at upper left indicate the RMSVE for M_2 (filled) and K_1 (empty) when DBDB2v30 topography is used. Plausible changes in topography result in tides which differ by much more than can be explained by uncertainty in C_d or in model resolution (cf., Table 1).

Figure 4: ETOPO1 and DBDB2v30 bottom topography.

Figure 5: a) ETOPO1 minus DBDB2v30 bottom topography. b) Bathymetric control points used in Smith and Sandwell topography v. 9.1, which slightly pre-dates ETOPO1. The red points are the locations of independent bathymetry (JAMSTEC cruise MR06-04) used to validate the topographic grids.

344 normalized by its maximum value (Figure 9) illustrates again the degree to
 345 which shallow, near-resonant, regions control the tidal elevation throughout
 346 the basin, even when the range of potential topographic perturbations is
 347 scaled with the local water depth. In this case, a 1% change in water depth
 348 at sites within Udkaya Guba (the dark-shaded regions in the western bay)
 349 has the same impact as a 30% change in water depth at the measurement
 350 site (light-green shading near the red star at mid-basin).

351 The non-log-scaled version of this field (Figure 10) illustrates how the
 352 sign of the topographic perturbation is related to changes in the in-phase
 353 component of η at the measurement site. These spatial patterns reflect the
 354 signs of the terms comprising λ in equation (18). Most of the sensitivity

Figure 6: Friction number, R_f . The dashed line, most visible near the northern coast, indicates the 100 m depth contour.

Figure 7: Terms balances, zoomed. a) R_f , friction number. b) Fr , Froude number. c) Z , vertical excursion number. Solid dashed line indicates the 100 m depth contour; gray dashed line indicates the 30 m depth contour.

355 results from the frictional term,

$$C_d u_f \frac{\boldsymbol{\mu}^{*(k)} \cdot \mathbf{U}^{(k)}}{H^2}, \quad (26)$$

356 with the magnitude of the term largely caused by the smallness of H . The
 357 geometry of the bays also leads to near-resonance and amplifies the adjoint
 358 $\boldsymbol{\mu}$ fields, too, just as in the LTE.

359 In comparison with the sensitivity to topography, the sensitivity to the
 360 drag coefficient is much smaller (Figure 11). The scaling of λ by H and c
 361 by C_d makes the numeric values directly comparable in the sense that same
 362 percentage perturbation in either H or C_d would yield a proportional change
 363 in the in-phase component of η at the measurement site. Because of the $1/H$
 364 scaling of c , compared to the $1/H^2$ scaling of λ , the sensitivity to C_d is limited
 365 to the near-coastal areas, and it is much smaller than the sensitivity to H .
 366 The sensitivity to changes in C_d may be regarded as measure of sensitivity
 367 to changes in the other frictional parameters, u_0 and γ_f , since these both
 368 multiply C_d in the expressions for the drag. Note that the dimensional values
 369 of $H\lambda$ and $C_d c$ shown are comparable to each other.

370 Other sensitivity fields have been computed corresponding to each of the
 371 270 measurement sites in Figure 1a and for measurements of both the in-
 372 phase and quadrature components of M_2 and K_1 , for 1080 distinct mea-
 373 surement operators, κ_i , in total. Perhaps the most striking feature of these
 374 sensitivity fields is their great visual similarity. Essentially all the sensitivity
 375 fields for η measurements in mid-basin are dominated by the near-resonant
 376 bays (not shown).

377 The similarity may be quantified by looking at the degree of linear in-
 378 dependence of the sensitivity fields, λ_i , resulting from different measure-

379 ments, $i = 1, \dots, 1080$. This has been done by computing the singular
 380 value decomposition of the matrix of inner products, $W = \int_{\mathcal{D}} \lambda_i \lambda_j$, and
 381 the depth-weighted version of this matrix, $W_H = \int_{\mathcal{D}} \lambda_i H^2 \lambda_j$, corresponding
 382 to the depth-weighted sensitivities shown above. The singular spectrum of
 383 W (Golub and Van Loan, 1989) indicates the linear dependence of the sen-
 384 sitivity fields, as exhibited by the rapid decay of the singular values with
 385 increasing index (Figure 12a). The singular spectrum of W_H decays more
 386 slowly, but even in this case, the decay over the first 10 singular modes is
 387 almost as fast as in the unweighted case. Another measure of the degree
 388 of linear dependence is the cumulative sum of the singular values compared
 389 to their total (the trace of the matrix). In the unweighted case, more than
 390 99% of the total sensitivity variance is contained in the first 10 modes (Fig-
 391 ure 12b); for the weighted sensitivities the first 10 modes contain about 90%
 392 of the sensitivity variance.

393 5. Discussion

394 The present work has quantified the sensitivity of tidal water elevations
 395 to bottom topography, friction, and other factors in the Sea of Okhotsk.

396 The motivation for this study was primarily to understand the difficul-
 397 ties with identification of bottom topography from measurements of tidal
 398 water elevation using variational data assimilation at continental-shelf scale.
 399 Experiments in the Sea of Okhotsk and in the South China Sea indicated
 400 that in realistic identical-twin type experiments, it is difficult to accurately
 401 reconstruct the bottom topography field from water level measurements (not
 402 shown). Detailed analysis of an idealized one-dimensional tide/wave model

Figure 8: Adjoint sensitivity. The logarithm of the absolute value of the adjoint sensitivity, λ , is shown for an η measurement near mid-basin indicated with the red star. The tidal elevation is most sensitive to the topography in the bays where tidal resonant amplification is large.

Figure 9: Adjoint sensitivity multiplied by H , log-scaled. The logarithm of the absolute value of $H\lambda$ is shown for the same measurement site as in Figure 8. The adjoint sensitivity is multiplied by depth to account for the size of depth changes in the sensitivity.

Figure 10: Adjoint sensitivity multiplied by H , non-log-scale.

Figure 11: Adjoint sensitivities, zoomed non-log-scale. The 100 m and 30 m depth contours are also shown.

Figure 12: Analysis of linear dependence of topographic sensitivity. (a) The spectra of singular values of the matrices $W = \int_{\mathcal{D}} \lambda_i \lambda_j$ (red) and $W_H = \int_{\mathcal{D}} \lambda_i H^2 \lambda_j$ (black) indicate the degree to which the adjoint sensitivity functions are linearly dependent. The controllability of the tidal surface elevation is determined by very few degrees of freedom of the bottom topography. (b) The cumulative sum of the singular spectra indicates that the first 10 degrees of freedom account for 99% (red, W) or 90% (black, W_H) of the η variance controlled by H , for the unweighted and weighted cases, respectively.

403 found that near-resonant continental shelves presented special difficulties
 404 because of the sensitivity to processes in the region of resonant amplifica-
 405 tion (Zaron, 2016). Because of the small spatial scale of these regions, they
 406 are necessarily under-observed compared to the remainder of the domain;
 407 and the resonant amplification is associating with poor conditioning and a
 408 reduction in the degrees of freedom which can be identified.

409 In the interests of space, the present study has investigated adjoint sen-
 410 sitivities in a realistic two-dimensional tide model without showing the com-
 411 plete inversion for topography. A thorough development of the topographic
 412 estimation problem would involve one additional element, namely, the spec-
 413 ification of a spatial error model for the topography. Such a development is
 414 lengthy and involves the consideration of multiple sources of bathymetry, and
 415 it is logically separate from the sensitivities discussed herein. Nonetheless,
 416 the scaling of the topographic sensitivity by H provides a more interpretable
 417 measure of sensitivity and is directly relevant to statistical models which
 418 parameterize the fractional uncertainty of H . If the difference in topogra-
 419 phy, $\Delta H = H_{ETOPO1} - H_{DBDB2v30}$, is regarded as a plausible estimate for the
 420 topographic error, one useful conclusion from the study of the degrees of free-
 421 dom of the W_H matrix is that the space spanned by the weighted sensitivity
 422 functions $\{H\lambda_i\}$ is very limited. For example, in the present case, project-
 423 ing ΔH onto the first forty eigenmodes of the W_H matrix only explains one
 424 percent of the variance of the ΔH field.

425 The results with this shelf-scale model stand in contrast to findings in
 426 riverine and nearshore ocean models, which have been used successfully to
 427 identify bottom topography from water level and other measurements. The

reason for this can be explained by the spatial locality of the adjoint sensitivities in the different cases. For example, in the riverine study of Zaron et al. (2011), the dominant dynamical balance was between the along-channel pressure gradient and the bottom stress. The along-channel pressure gradient was determined independently, and, therefore, there was essentially a one-to-one relationship between the observed data (near-surface current) and the bottom depth at the observation location. Thus, each current measurement provided an independent constraint on bottom depth local to the measurement site. Likewise, much of the work on nearshore models, e.g., Wilson et al. (2014), involves the assimilation of wave celerity or other quantities which provide spatially localized constraints on the water depth. Unlike these cases, for tidal dynamics in the Sea of Okhotsk, each water elevation measurement is sensitive to topography at a relatively small number of locations; the data provide non-independent observations of a small set of topographic degrees of freedom.

To what extent are these findings generalizable to settings outside the Sea of Okhotsk? The variational approach to the identification of topography requires both tidal elevation data and a dynamical model which are relatively more accurate than the bottom topography. In the deep ocean there is obviously no shelf-scale resonant amplification of tides, so the results are not relevant there; however, the fractional uncertainty of topography is much smaller than the dynamical or data uncertainties, so the prospect of improving the topography there is a moot point. It is on the continental shelves where bathymetry is sparse and where the gravimetric topography is inaccurate that one might wish to apply the variational approach; but small-

453 scale resonant amplification is not uncommon near the coast, and it appears
454 that the variational methodology is fundamentally flawed for identifying to-
455 pography in most cases of practical interest.

456 In contrast to the negative outlook for identification of topography, the
457 present results suggest the potential for monitoring coastal processes using
458 offshore altimeter measurements. For example, the dissipation of tidal cur-
459 rents near the edges of some Antarctic floating ice shelves is apparently a
460 sensitive function of the ice shelf thickness (Mueller, 2014). The latitude
461 of the ice shelves is poleward of the region sampled by the highly-accurate
462 TOPEX/POSEIDON-Jason series of satellite altimeters; however, it may be
463 possible to use these more distant altimeter data to either hindcast or mon-
464 itor ice shelf thickness or extent.

465 The present results are consistent with the findings of other studies that
466 indicate the tidal fields can be optimized treating the bottom topography as
467 a control parameter (e.g., Ngodock et al. 2016). In that case, the goal is to
468 improve η or other derived quantities in the tide model, rather than making
469 verifiably improved estimates of the bottom topography per se. The lack of
470 linear dependence of the adjoint sensitivity functions suggests that a small
471 number of degrees of freedom are necessary to effect significant changes in
472 the tides by adjusting the topography.

473 **6. Summary**

474 The sensitivity of barotropic tides to bottom topography and frictional
475 parameters has been studied in a model for the Sea of Okhotsk. This region
476 was chosen because of the paucity of bathymetry data in the region and

477 the possibility of using satellite altimeter data to better identify the bottom
478 topography using variational inverse methods. Sensitivity to topography and
479 other factors was studied using both the direct and adjoint sensitivity. In the
480 former approach, various perturbations to the nominal model were applied to
481 examine their impact; in the latter approach, the sensitivities were computed
482 using the adjoint of the tangent linearization of the dynamical model.

483 Direct comparisons of the tide models was conducted at different spatial
484 resolutions, using different drag coefficients, and using different topographies.
485 Amongst these factors it was found that plausible changes in topography pro-
486 vided the largest impact on the tidal elevations, as measured by comparison
487 with tides observed by satellite altimetry. Although the model’s realism is
488 limited by the omission of nonlinear terms, flooding and drying processes,
489 and the finite number of tidal frequencies used, the influence of these factors
490 appears to be smaller than either the topography or frictional parameteriza-
491 tions.

492 The adjoint sensitivities computed with the model clarify the roles of both
493 topography and friction in determining the tidal elevation. The principal
494 finding is the extent to which small-scale near-resonant amplification controls
495 tides at the basin scale. This is not a new finding (Lyard and Genco, 1994;
496 Lefevre et al., 2000); instead, the results clarify the relatively few degrees of
497 freedom in the topography which control the tides. The nature of the control
498 depends on both the geometry of the basin, which permits the small-scale
499 amplification, and also the presence of a bottom dissipation, which amplifies
500 the sensitivity in shallow water by the factor $1/H^2$.

501 The findings reiterate the conclusions of Zaron (2016) regarding the signif-

502 icance of small-scale near-resonant amplification for the topographic identifi-
503 cation problem using tidal dynamics. In contrast to results with riverine and
504 nearshore ocean models, where the observations are more closely related to
505 topographic features localized to the measurement location, the tidal dynam-
506 ics and resonance creates a non-local relationship and lack of independence
507 of the data. As in the idealized one-dimensional study of Zaron (2016), it
508 is hypothesized that the topographic identification problem is sensitive to
509 the details of the spatial covariance model for the topography in regions of
510 tidal amplification. This fundamentally limits the applicability of inverse
511 approaches for identification of topography, and provides an additional ra-
512 tionale for the collection and sharing of high quality bathymetry data to
513 enable improved ocean modeling.

514 **7. Acknowledgments**

515 Support from the NASA Ocean Surface Topography Science Team award
516 #NNX13AH06G is gratefully acknowledged. Satellite altimeter data used
517 here were extracted from the Radar Altimeter Database System (RADS;
518 <http://rads.tudelft.nl/rads/rads.shtml>), an invaluable resource. The
519 software for this project was derived from the Oregon State Tidal Inversion
520 Software (OTIS; <http://volkov.oce.orst.edu/tides/otis.html>), devel-
521 oped and kindly provided by Gary Egbert and Lana Erofeeva.

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