1 Introduction

Oceanic gravity waves fundamentally contribute to the redistribution of mechanical energy in the world ocean. The wave transmission can be altered as a result of interaction with large scale sea bottom features such as submarine mountains, trenches and continental slopes. This redistributes energy, causes wave reflection or might transform wave structure. In this work, by means of numerical simulations, these processes are examined for tsunamis and internal waves of tidal frequency (internal tides) for realistic cases of their interaction with ocean's bottom relief.

Tsunamis are transient waves that are impulsively forced by submarine earthquakes. The waves traverse vast expanses of the oceans and encounter numerous sea bottom features (Mofjeld et al., 2001). In each instance, scattering redirects portion of incident energy. This process in case of the Hawaiian-Emperor seamount chain and Hess Rise leads to highly variable spatio-temporal tsunami impact on coastal regions (Kowalik et al., 2008; Tang et al., 2012). First, this is characterized by wave trains arriving in irregular sequence. The largest waves can unexpectedly appear with a several hour delay after the initial arrival (Koshimura et al., 2008). At second, the overall wave intensity can markedly change along shoreline (Borrero & Greer, 2013). The problem is investigated in the Chapter 1. Scattering by Koko Guyot, the largest submarine seamount in the Emperor chain (Figure 1), is studied in order to quantify for the observed far-field signals in two recent tsunami events: the 2006 Kuril and 2011 Tohoku-oki tsunamis.

The second problem considered in this dissertation explores internal waves that pervade ocean's interior. Their existence depends upon water column stratification since these waves manifest as an oscillatory motion of isopycnal surfaces. The initial perturbation can be forced by different mechanisms: wind stress (Garrett, 2001), buoyancy flux (Garrett & Munk, 1979), ice ridge keels (Morison, 1986) or even tsunami (Santek & Winguth, 2007). Additionally, tidal currents at steep topography can excite internal waves. These are baroclinic waves that oscillate with tidal and are called internal or baroclinic tides.

One reason to study baroclinic tides is their role in the dissipation of barotropic (surface) tidal energy (Munk, 1997). Current estimates suggest that 1/3 of total tidal energy is converted into internal motions in the deep ocean (Egbert & Ray, 2000). From this roughly 1/3 is directly lost to turbulent mixing right at generation sites (St. Laurent & Garrett, 2002) while the rest is emitted as freely propagating waves in form of tight beams (Simmons et al., 2004). These structures are observed to cross the ocean basins without much attenuation (Zhao et al., 2016). This finding fuels scientific research to understand where and how the baroclinic tidal energy is dissipated. Since the deposition sites primarily would occur deep below direct wind influence, the internal tides are thought to be a candidate for providing mechanical energy to close the upwelling branch of the Meridional overturning circulation (Munk & Wunsch, 1998). Regardless of global importance, the baroclinic tides are relevant for local processes such as water mass transformation (Stigebrandt & Aure, 1989), vertical nutrient fluxes (Sharples et al., 2007), sediment (Hotchkiss & Wunsch, 1982) and larvae transports (Pineda, 1999). All of the above is crucially dependent on an efficiency of energy removal in the process of internal-tide generation, propagation and scattering. And the respective parameterization is in a premature state to have a practical use.

In the setting of the Tasman Sea, an internal tidal beam originates at Macquarie Ridge and impinges on the continental slope of Tasmania (Figure 1). The Tidal Dissipation Experiment (TTIDE) had its goal to provide description of the internal tide reflection and involved energy lost. This means to quantify amount of wave energy being reflected back into open sea, converted into higher vertical modes and consequently, describe processes associated with sink of energy. These are the problems addressed in Chapters 2 and 3.

The two geophysical problems examined in the dissertation represent wave scattering by inhomogeneity in propagation media. Here, it takes form of rapid change in ocean depth imposing constrains on fluid motion. The accompanying transfers are intrinsically linked to energy flux approach. This is reviewed in **Section 2** of the Introduction by considering a normal mode dynamics, while **Section 3** describes energy characteristics associated with wave transformations caused by interaction with bathymetry.

2 Normal modes in a flat-bottom sea

Tsunami and internal tides are long gravity waves. Their typical wavelengths are order of 100 km being larger than the ocean's mean depth. Thus, the wave dynamics can be well described under the linear, hydrostatic framework. As a first step, wave propagation in a flat-bottom ocean is discussed. The equations of motion in stratified, Boussinesq ocean on an f-plane (Kundu & Cohen, 2008; Cushman-Roisin & Beckers, 2011) will take mathematical form of

$$\frac{\partial u}{\partial t} - fv = -\frac{1}{\bar{\rho}_0}\frac{\partial p}{\partial x} \tag{1a}$$

$$\frac{\partial v}{\partial t} + fu = -\frac{1}{\bar{\rho}_0}\frac{\partial p}{\partial y} \tag{1b}$$

$$0 = -\frac{\partial p}{\partial z} - \rho g \tag{1c}$$

$$N^2 w = -\frac{1}{\bar{\rho}_0}\frac{\partial^2 p}{\partial z \partial t} \tag{1d}$$

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0 \tag{1e}$$

where $(u,\ v,\ w)$ are velocity along zonal, meridional and vertical axis, f is the Coriolis parameter, p - perturbation pressure arising either from sea level oscillations or isopycnal displacements within stratified water column measured by Brunt-Vaisala frequency $N^2 = -\frac{g}{\bar{\rho}_0}\frac{\partial \rho_0}{\partial z}$. Additionally, boundary conditions are imposed on the surface and bottom,

$$w|_{z=0} = \frac{\partial \zeta}{\partial t},\ p|_{z=0} = \bar{\rho}_0 g \zeta \tag{2}$$

$$w|_{z=-H} = 0 \tag{3}$$

Both conditions are of linear form since the sea level motions are much smaller than the acceleration due to gravity, and the ocean depth is fixed. This formulation allows a solution via separation of vertical motions by employing following decomposition,

$$(u,v,p)(x,y,z,t) = \sum_{n=0}[u_n(x,y,t), v_n(x,y,t), p_n(x,y,t)]\psi_n(z) \tag{4}$$

$$w(x,y,z,t) = \sum_{n=0}[w_n(x,y,t)]\int_{-H}^{z}\psi_n(z)dz \tag{5}$$

$$\rho(x,y,z,t) = \sum_{n=0}[\rho_n(x,y,t)]\frac{d\psi_n(z)}{dz} \tag{6}$$

Upon substitution into (1) Sturm-Liouville problem is obtained for vertical structure (basis) functions $\psi_n(z)$[1],

$$\frac{d}{dz}\left(\frac{1}{N^2}\frac{d\psi_n}{dz}\right) + \frac{1}{c_n^2}\psi_n = 0 \tag{7}$$

[1]note this is for horizontal variables, usually it is written for vertical velocity since boundary conditions are expressed right away

with an eigenvalue c_n for a respective n-th vertical mode. c_n is the wave phase speed in a non-rotating ocean with an equivalent depth, D_n such that $\sqrt{gD_n} = c_n$. The decomposition separates vertical motions from horizontal dynamics reducing the initial set (1) to the Laplace tidal equations,

$$\frac{\partial \vec{u}_n}{\partial t} + f\vec{k} \times \vec{u}_n = -\frac{1}{\rho_0}\nabla p_n \tag{8a}$$

$$\frac{1}{\rho_0}\frac{\partial p_n}{\partial t} + gD_n \nabla \cdot \vec{u}_n = 0 \tag{8b}$$

This reveals the dynamical equivalence between barotropic, surface wave and baroclinic, internal waves if the ocean stratification were absent and the depth was D_n. In actuality, 0-th mode represents a barotropic solution with vertically uniform dynamics ($\psi_0 = const$) and the depth $D_0 \simeq H$ (Hendershott, 1981). The gravest mode represents pure surface wave propagation such as in the tsunami wave problem. The infinite sequence of higher vertical modes are internal modes that produce increasingly negligible sea level perturbations. This allows complete separation between barotropic and baroclinic motions which is similar to imposing rigid lid condition on internal modes (Kundu & Cohen, 2008).
The main interest is in linear wave phenomena, so harmonic temporal behavior, $e^{i\omega t}$ will be implied leading to the Helmholtz equation,

$$(\omega^2 - f^2)p_n + c_n^2 \Delta p_n = 0 \tag{9}$$

Making a plane wave ansatz, $e^{-i\vec{k}\cdot\vec{x}}$, currents and pressure will be related as

$$\begin{aligned} p_n &= p_{0n} e^{i(\omega t - \vec{k}\cdot\vec{x})} \\ u_n &= \frac{1}{\bar{\rho}_0}\frac{\omega k_x + ifk_y}{\omega^2 - f^2} p_n \\ v_n &= \frac{1}{\bar{\rho}_0}\frac{\omega k_y - ifk_x}{\omega^2 - f^2} p_n \end{aligned} \tag{10}$$

Then from (8b), a dispersion relation for Sverdrup waves follows,

$$\omega^2 = f^2 + c_n^2 |\vec{k}|^2 \tag{11}$$

The waves are dispersive as a result of Earth's rotation, with phase and group speeds are

$$c_p = (1 - \frac{f^2}{\omega^2})^{-1/2} c_n, \;\; c_g = (1 - \frac{f^2}{\omega^2})^{1/2} c_n, \;\; c_p c_g = c_n^2 \tag{12}$$

For the most energetic mode-1 semidiurnal internal tide and typical ocean conditions phase travels with speed of 4 m/s, but group speed is only 2 m/s. So slow traveling wave is potentially affected by background oceanic circulation. On contrary, tsunami waves exhibit shallow-water, nondispersive behavior with phase and group speeds of $\sim O(100\ m/s)$ owing to small periods in comparison to inertial period. Tsunami waves cross ocean basins in a matter of hours without any appreciable interaction with ocean's dynamical state.

In the deep ocean, where the bottom topography does not demonstrate large spatial gradients, normal modes are useful to describe wave dynamics. An energy equation can be developed from (8). Multiplying the momentum equations by $\vec{u}_n$ and the continuity equation by p_n, adding both expressions and depth integrating one obtains

$$(E_k + E_p)_t + \nabla \cdot \vec{F} = 0 \tag{13}$$

Additionally, depth averaging is carried out so that energy flux will be defined,

$$\vec{F}_n = \frac{1}{H}\vec{u}_n(t)p_n(t)\int_{-H}^{0}\psi^2(z)dz \tag{14}$$

For tsunami wave propagation due to their barotropic nature $\psi(z) = const$ and $p_n = \rho_0 g\eta$, the expression takes its classical form for surface wave propagation (Henry & Foreman, 2001), $\vec{F} = \rho_0 g\vec{u}\zeta$.
In internal tide studies period averaging further simplifies calculations,

$$\vec{F}_n = \frac{1}{2H}\vec{u}_n^{\star}p_n\int_{-H}^{0}\psi^2(z)dz \tag{15}$$

where velocity and baroclinic pressure are now complex amplitudes of harmonic fit and * is a complex conjugate.
The obtained expressions for energy flux comprises only the pressure-work term, i.e. transport of energy is dictated by pressure forces. In full formulation this is augmented with nonlinear contributions made by advection and dissipative forces (Gill, 2016). These terms are negligible presuming deep ocean propagation where the wave's spatial dimensions are large and magnitude of dynamical variables is small.
Away from scattering regions, an energy budget can be constructed following the linearized normal mode dynamics. Any loss in intensity of primary wave has to be balanced by energy scattered into other form of motion such as vertical modes or trapped waves and by dissipative losses. The latter is not discussed here and thought to constitute a residue term. The scattered wave field, reflected components and intermodal conversion are the main topic of investigation. The quantitative aspects are discussed in **Section 3**.

3 Oceanic wave scattering

The normal mode approach was formulated under the assumption of flat bottom with a simplified boundary condition, $w|_{z=-H} = 0$ that allows separation of variables. Thence, vertical motions are decoupled from horizontal dynamics. As a wave encounters an inclined slope the simplified relation does not hold and the bottom boundary condition becomes nonlinear,

$$(\vec{u} \cdot \nabla h)|_{z=-h(x,y)} = 0 \text{ or } w = -\vec{u}_h \cdot \nabla h \tag{16}$$

The condition states that there is no mass transport across an impermeable bottom. The wave motion cannot alone satisfy (16), and a compensating flow has to develop to preserve

mass. For the configuration when an incident wave train encounters a vertical wall in nonrotating ocean, a specular reflection occurs in order to fulfill the no-flow condition. Opposing fluid motions will be produced by a back-scattered wave.
An additional condition is imposed on scattered waves such that the energy vanishes with distance (Mei et al., 1989; Morse & Feshbach, 1946). The Sommerfeld radiation condition leads to a far-field asymptotic:

$$p_{scat} \sim \frac{e^{i\vec{k}\cdot\vec{x}}}{|\vec{x}|}\Phi_s(\theta) \tag{17}$$

3.1 Tsunami wave scattering

This is an assertion for circularly spreading waves having some angular distribution Φ_s, called a scattering amplitude. The squared quantity will furnish amount of scattered radiation. In the problem of tsunami-topography interaction, far-field signals could be approximated with the above distributions; while in internal tide reflection problem, large distances away (when continental slope becomes a point) an angularly integrated scattered amplitude will be directly related to a reflection coefficient, though wave spreading with distance has to be factored in.
In the near-field, a scattered wave field can be decomposed by Fourier transform whereby each wavenumber can be regarded as a separate component. As it propagates away, it will begin to separate from others. Magnitude modulations due to interference [2] will subdue. And at large distances, respective Fourier coefficient will represent a scattering amplitude in direction corresponding to the wave vector. As a result, Fourier transform depicts the far-field's distribution of energy.[3]

3.2 Internal tide scattering

In addition to angular scattering, there could occur a transfer of energy into different vertical modes. Let us consider a long wave that obeys Laplace tidal equations entering a region where bottom slope is inclined and the full boundary condition (16) has to be satisfied. It only can be fulfilled if there is a vertical velocity. A normal mode by itself is not able to represent the dynamics since the vertical derivative, by definition, is zero on the boundary. However, a superposition of modes can force horizontal convergence to cancel out the vertical velocity. Physically, pressure gradients should develop by displacing isopycnal surfaces which further radiate energy away in the form of baroclinic modes. In general, this describes a coupled-mode approach (Griffiths & Grimshaw, 2007). Locally-flat normal modes counteract one another through the boundary condition. The spatial gradients will dictate mode's magnitude to equilibrate the balance. Dynamically, Laplace tidal equation for each separate mode will embrace an additional force represented as intermodal-coupling term (Griffiths & Grimshaw, 2007; Kelly et al., 2016).
The intermodal conversion and energy flux carried by modes can be estimated as work

[2] more appropriate to say diffraction
[3] This is a fundamental idea in acoustic holography (Williams, 1999), should I cite other fields?.

done by vertical velocity against modal pressure, $W_{m\to n} \sim (w_{m,\ scat}\ p_n)|_{z=-h} = -((\vec{u}_m \cdot \nabla h)p_n)|_{z=-h}$. But total conversion must as well account for the opposite energy transfer, from n to m, hence,

$$C_{m\to n} = -((\vec{u}_m \cdot \nabla h)\ p_n)|_{z=-h} + ((\vec{u}_n \cdot \nabla h)\ p_m)|_{z=-h} \tag{18}$$

Internal tide excitation can be viewed as scattering of barotropic tide by steep relief (Hendershott, 1981). If the surface tide is the 0^{th} mode, then energy lost can be diagnosed as

$$C_{bt\to bc} = -(\vec{u}_{bt} \cdot \nabla h) \sum_{n=1} p_n|_{z=-h} \tag{19}$$

assuming that the baroclinic wave field effect on surface tide is negligible (Kelly et al., 2012). The above expression is well established in internal tide studies (Kurapov et al., 2003; Llewellyn Smith & Young, 2002; Pickering et al., 2015).
For consideration of mode-1 scattering it is useful to reformulate (18) by employing Leibniz's rule, the continuity equation per each normal mode, and mode's orthogonality,

$$C_{m\to n} = \int_{-h}^{0} (u_m \nabla p_n - u_n \nabla p_m) dz \tag{20}$$

In the problem of internal wave propagation over uniform and gently inclined bottom (Wunsch, 1968), there is no intermodal conversion, $C_{m\to n} = 0$, and each mode's dynamics is perfectly matched by presence of others. Strictly speaking, in this case the wave equation is hyperbolic [4] and propagation occurs along characteristics (Sandstrom, 1969) or internal wave rays described by

$$\gamma = \frac{dz}{dx} = \pm \frac{\omega^2 - f^2}{N^2 - \omega^2} \tag{21}$$

Slope criticality can be defined as ratio between its inclination and rays angle, γ. In subcritical regime of forward propagation, the ratio is less than unity. In this case for steeper bottom more higher modes is necessary to describe internal tide dynamics (Figure 2). Now it is apparent that a mode conversion will not happen, unless there is a change in topography to drive adjustment of mode amplitudes. On contrary, in supercritical regime, when $\nabla h > \gamma$, internal waves are reflected with always present generation of higher modes. Usually, they will be phase-locked into vertically propagating rays (Garrett & Kunze, 2007; Lamb, 2014) subject to breaking and dissipation (Lien & Gregg, 2001; Nash et al., 2004; Klymak et al., 2011). In case of the continental slope this causes internal tide mode-1 reflection to decrease relative to the equivalent case of surface wave reflection.
As a first approximation for long waves scattering, it will be instructive to consider a problem with simplified geometry. This will usually consist of a step discontinuity connecting the deep ocean with the top of seamount (e.g. Koko Guyot) or the continental shelf. Then the boundary condition (16) is analogous to fulfilling so-called matching conditions (Mei et al., 1989). These are mass conservation and continuity of sea level for surface waves

$$\vec{u}_1 h_1 = \vec{u}_2 h_2,\ \zeta_1 = \zeta_2 \tag{22}$$

[4](in contrary to elliptical Helmholtz equation, (9))

In the case of internal tides - continuity of pressure and no-flow through vertical walls (St. Laurent & Garrett, 2002; Larsen, 1969; Chapman & Hendershott, 1981), the conditions are

$$p_1 = p_2, -h_2 \leq z \leq 0 \tag{23}$$

$$\vec{u}_1 = 0, -h_1 \leq z \leq -h_2 \tag{24}$$

$$\vec{u}_1 = \vec{u}_2, -h_2 \leq z \leq 0 \tag{25}$$

4 Organization of thesis

The thesis consists of the following three chapters:

Chapter 1 makes its goal analysis of scattering of tsunami waves by Koko Guyot. The approach is based on obtaining scattering amplitude (17), evaluation of its frequency and spatial characteristics in order to describe far-field signals. The discussion and conclusions are supported with analytical calculations based on matching conditions approach.

Chapter 2 deals with generation of semidiurnal internal tide at Macquarie Ridge, Tasman Sea. This is studied in a detail by analysis of three dimensional structure of internal wave field. The proposed mechanism explains peculiarities of conversion rates (19). The emitted internal tidal beam than described in the far-field with emphasis on its kinematic properties. **Simple statement: generation of internal tides is affected by the remote signal -¿ this progresses into far-field.**

Chapter 3 concerns the tidal beam reflection from Tasman Continental Slope. The obtained energy levels show variability that is due to number of factors. Possibility of each is discussed and respective estimates are made in order to propose physically reasonable mechanism of internal tide dynamics in three dimensionally complex region.

Throughout the thesis a method based on inverse modeling was used. Its mathematical rudiments are given in Appendix A.

Figures

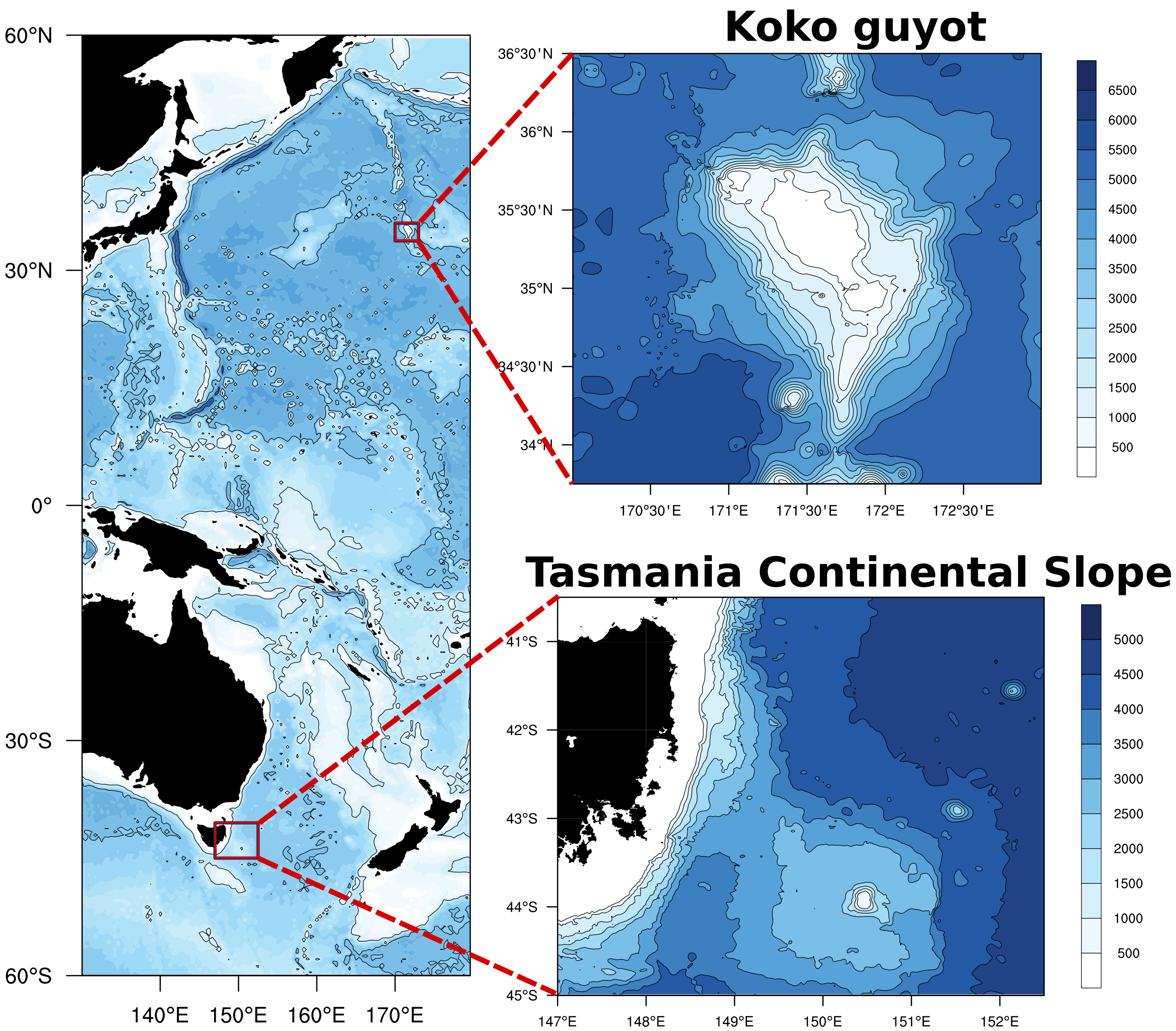

Figure 1: Bathymetric map of the Pacific Ocean with outlined regions showing geographic locations of considered wave scattering problems.

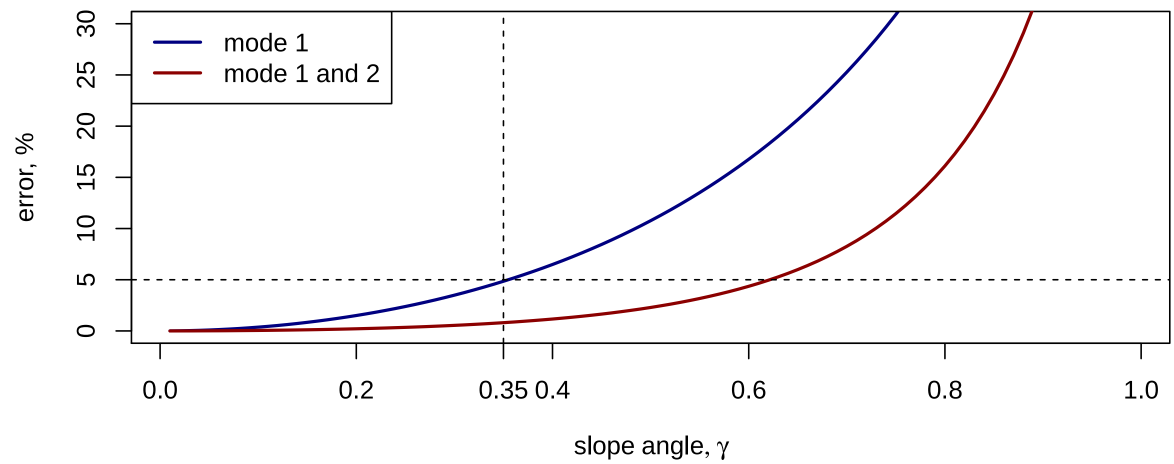

Figure 2: Error in description of internal wave structure by flat-bottom vertical modes as bottom gets steeper.

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