THE SEA: THE SCIENCE OF OCEAN PREDICTION
Journal of Marine Research, 75, 239–262, 2017

Advanced wave modeling, including wave-current interaction

by Alexander V. Babanin[1,2], Andre van der Weshuijsen[3], Dmitry Chalikov[4], and W. Erick Rogers[5]

ABSTRACT

The paper outlines principles of phase-resolving and phase-average wave models, with emphasis on the state of the art of wave-current interaction physics. We argue that these interactions are the least well-developed part of such models. Linear and nonlinear dynamics of waves on currents are discussed; depth-integrated and depth-varying approaches are described. Finally, examples of numerical model performance for waves on currents in realistic oceanic scenarios are presented.

Keywords: phase-resolving models, spectral models, wave-current interactions, wave modeling

1. Introduction

By the beginning of the new century, three major technological advances brought the oceanographic science essentially into a new era. These are satellite remote sensing, Argo fleet, and modern computer modeling. The former two provide the first ever opportunity of global long-term observations of the ocean surface properties—including winds, waves, and, to a lesser extent, currents—and three-dimensional ocean dynamics, respectively. The numerical modeling has taken on a basically new function. Provided the physics of the models is well defined and the limits are well understood, it can play both a role of a virtual laboratory or even of the ocean, with unlimited number of sensors and probes in every location in space and at any moment in time, and of an analytical tool to scrutinize known features and discover new phenomena. In case of the ocean and its waves, the models can also be used for forecast: to predict the future conditions, and for hindcast: to study the past ocean states.

1. The University of Melbourne, Parkville, Victoria 3010, Australia
2. Corresponding author: ***e-mail: a.babanin@unimelb.edu.au***
3. NOAA, National Weather Service, National Centers for Environmental Prediction, Environmental Modeling Center, College Park, Maryland 20740, USA
4. St. Petersburg Branch, P. P. Shirshov Institute of Oceanology, RAS, Saint Petersburg, Russia
5. United States Naval Research Laboratory, Washington, D.C. 20375, USA

Ocean waves, even in simple, deep-water background conditions, represent one of the most challenging research objects from the experimental, analytical, modeling, and statistical points of view (see, e.g., Phillips 1966; Young 1999). Generated and forced by the turbulent wind at the scale of thousands of wave periods (Janssen 2004), ocean waves are subject to dissipation and strong nonlinear exchanges at an intermittent scale of tens of wave periods due to wave-breaking events that last a fraction of a wave period (Babanin 2011), complemented by weakly nonlinear interactions that drive the evolution of wave spectra at the scale of tens of thousands of wave periods (Hasselmann 1962; Zakharov 1968). Ocean waves are superposed by a variety of less common or weaker processes, such as rogue waves linked to modulational instability of wave trains (e.g., Kharif et al. 2009; Babanin and Rogers 2014), or upper-ocean mixing due to wave-induced turbulence (Kinsman 1965; Ghantous and Babanin 2014), which are of great significance in some circumstances and applications. It should be stressed that most wave flumes, analytical theories, and numerical models are two-dimensional (2D), whereas the ocean is three-dimensional (3D), and many wave processes in the three dimensions are different or altered significantly (e.g., Chalikov et al. 2014).

Wave modeling efforts and applications can be broadly classified into two large groups: phase-resolving (or direct) models and phase-average (usually spectral) models. Direct models can explicitly simulate basic equations of fluid mechanics for the water, air, or even two-phase media, and therefore extend the analytical research beyond its traditional range of approximate and asymptotic solutions of such equations.

At oceanic scales, such models are not practical, and thus spectral models are employed for wind-wave forecast. The spectral models are based on parametric, often speculative approaches, but their physics lately has been advancing (see, e.g., Lavrenov 2003; Cavaleri et al. 2007).

It should be mentioned that operational wave forecast is much older than the oceanographic forecast. It is usually accepted as having started in 1942 for military needs (Rogers et al. 2014), but one may go back in time to find earlier mention (Gain 1918; Montagne 1922). Ideas of predicting the wave spectrum can be traced back to Gelci et al. (1956, 1957), and since then the second- and third-generation wave models have been progressing alongside technical developments in computing facilities. By comparison, the ocean forecast is only about 20 years old (but not the tide predictions, which are much older). A side product of such an early start is that for decades the wave forecast was considered a part of meteorology. In a way, it can be justified since the waves are wind-generated, but the surface waves are mostly part of the motion on the water side of interface and are strongly linked to the upper-ocean dynamics including the ocean currents. It is only recently that wave modeling has begun to establish its place in the mainstream oceanography. In this review, the oceanographic role of surface waves is discussed.

In finite-depth and shallow environments, waves essentially become a different physical object (see, e.g., Holthuijsen 2007). Dispersion is reduced, or even ceases, nonlinearity grows, but active nonlinear mechanisms change, the balance between energy input and dissipation is no longer maintained, and a variety of new physical processes come into existence because of various wave-bottom interactions and sediment response. Respective wave models are notable for a lesser degree of physics and greater degree of parametric and ad hoc tuning.

While parameterizations of white-capping dissipation and wind input has progressed largely over the last few years, the most uncertain physics for ocean wave modeling and forecast currently rests on wave-current and wave-ice interactions, when and where such specific conditions occur. Wave-ice interactions have long been an exotic field of research, but with the Arctic opening from ice in summer months, wave-ice modeling acquires important practical meaning. Among the various theories to explain wave-ice interactions, some differ qualitatively, i.e., wave scattering (without dissipation) and dissipation (with or without scattering); others differ quantitatively to the extent that some theories predict wavelength to increase in presence of ice, whereas others predict wavelength to decrease. In the field, all the mechanisms are acting together, depending on their relative magnitude, and practical guidance of the existing theoretical knowledge in forecasting waves in marginal ice zones is limited. Additional complications in this regard are due to necessity of also knowing initial conditions for the ice coverage and properties, and to be able to predict effects of waves on ice—this makes wave-ice interaction an essentially coupled problem.

As far as ocean currents are concerned, these are common conditions both in the open ocean and in coastal areas. Major currents such as the Gulfstream, Kuroshio, or Agulhas are well known for harsh seas and high likelihood of abnormal (rogue) waves. Tidal inlets with waves on strong and variable currents are a typical feature of shipping routes in coastal areas. While linear effects of currents on waves, such as refraction, Doppler shift, or relative speed with respect to the wind are assumed to be implicitly or explicitly included in wave-forecast models (often unverified and not validated), nonlinear effects are usually left out or even unknown. These include changes to nonlinear interactions in the presence of currents with horizontal or vertical velocity gradients, wave–current energy and momentum exchanges, nonlinear modifications of the wave spectrum.

The last, but not the least, modeling waves in the context of wave-induced effects in the lower atmosphere and the upper ocean should be mentioned (e.g., Babanin et al. 2012). It is rapidly becoming clear that many large-scale geophysical processes are essentially coupled with the surface waves, and those include weather, tropical cyclones, storm surges, climate, and other phenomena in the atmosphere, at air-sea and sea-land interfaces, and many issues of the upper-ocean mixing and ocean currents below the surface. Besides, the wind-wave climate itself experiences large-scale trends and fluctuations (Young et al. 2011), and is subject to changes in the weather climate.

2. Phase-resolving models

Phase-resolving modeling of surface waves is the mathematical modeling that includes explicit reproduction of the sea surface and velocity field evolution. Compared with spectral-wave modeling, phase-resolving modeling is more general since it is closer to reality and operates with basic equations and a larger volume of variables. Generally, therefore, this method is more complicated and takes more computational resources. The complexity of wave modeling depends on additional assumptions. Periodic models usually use the precise Fourier Transform method, so they are well suited for long-term numerical simulations of wave field transformation due to nonlinearity, input, and dissipation of energy. For nonperiodic processes, such as waves in a restricted domain or waves above nonperiodic bottom topography, the Fourier presentation is inapplicable and numerical schemes should be based on finite-difference presentations. Usually, such models are not as exact as the models based on Fourier presentation, but in most cases the nonperiodic approaches consider short-lasting processes when very high accuracy is not necessary.

The phase-resolving models are based on the equation of potential motion with free surface: the Laplace equation for velocity potential and two boundary conditions - the kinematic condition on surface and the Bernoulli integral (dynamic condition). The simplest way of phase-resolving modeling is calculation of wave field evolution on the basis of linear equations. Such approach allows us to reproduce the main effects of the linear wave transformation due to superposition of wave modes, reflections, refractions, among others (e.g., Belibassakis 2007). This approach is useful in many technical applications, but obviously cannot be employed when it is needed to reproduce the nonlinear nature of waves and transformation of a wave field due to nonlinearity.

The nonlinearity can be taken into account in more sophisticated models that are derived from the fundamental fluid mechanics equations with or without some simplifications. The most popular approach is based on the nonlinear Schrödinger equation of different orders (see Dysthe 1979), obtained by expansion of surface wave displacement. This approach was very useful for many applications including the problem of rogue waves. The main advantage of the simplified approach is that it allows us to reduce the 3D problem to the 2D (or 2D to the 1D problem). However, it is not always clear which nonrealistic effects are eliminated or included in the model after simplifications.

This is why the most general approach being developed over the past few years is based on the initial 3D or 2D equations (still potential). As mentioned above, all tasks based on these equations can be divided into two groups: the periodic and nonperiodic problems. Assumption of periodicity considerably simplifies construction of numerical models, but such formulation can be applied only when domain is considered as a small part of a large uniform area. For the limited domains with no periodicity, the problem becomes more complicated.

Standard method of solution in a 2D case is based on introducing the conformal coordinates and reducing the problem to the one-dimensional. This allows us to make very simple and fast calculations of wave dynamics (Chalikov and Sheinin 1998). The scheme was found to be very precise, i.e., typical accuracy of solution for a sufficiently high resolution was around 10^{-10}. Such accuracy is not surprising, since the equations written in the conformal coordinates become 1D evolutionary equations that can be accurately solved by means of the Fourier Transform method without local approximations. High accuracy of solution and conservation of integral invariants are both crucial for the numerical wave simulations. The ratio of timescale for waves (wave period) to timescale for the energy input and dissipation is of the order of 10^{-3}. Hence, wave motion is highly conservative, i.e., at time scales of the order of wave period (in the absence of breaking), the motion is actually adiabatic.

The equations derived give a unique example of a fully nonlinear description of a natural process with a nearly computer accuracy. This approach can be applied for investigation of wave dynamics when the two-dimensionality of spectrum does not play an essential role, for example, for investigation of rogue wave dynamics. The 1D approach, however, considers an idealized wave field, since even the monochromatic waves in the presence of lateral disturbances quickly develop a 2D structure for their crests (e.g., Pinho and Babanin 2015). A difficulty thus arising is not a direct result of increase of the dimension. The principal complication is that, in three dimensions, the problem cannot be reduced to the two-dimensional, and even for the case of a double-periodic wave field, there remains the problem of solution of the Laplace equation for the velocity potential. Effective methods in this regard are based on the full 3D equations and surface integral formulations (Clamond and Grue 2001). The main advantage of the Boundary Integral Equation Methods (BIEM) is that they are accurate and can describe highly nonlinear waves. The method of solution of the Laplace equation is based on use of the Green function, which allows us to reduce a 3D water-wave problem to a 2D boundary integral problem.

The surface integral method is well suited for simulations of wave effects due to very large steepness, and specifically for investigation of rogue wave generation and wave breaking. However, the BIEM method seems to be very complicated and time consuming if applied to a long-term evolution of a multi-mode wave field in large domains. Therefore, applications of the method were illustrated by simulations of relatively simple wave fields, and it is unlikely that the method can be applied to the simulation of long-term evolution or a large-scale wave field with broad spectrum. Implementation of the multi-pole technique for a general problem of sea wave simulations solves this problem, but obviously leads to considerable algorithmic difficulties.

Currently, the most popular approach is the High-Order Spectral Model (HOSM) developed by Dommermuth and Yue (1987) and West et al. (1987). The method is based on standard presentation of the Laplace equation solution in Cartesian coordinate system and extrapolation of this solution onto free surface via the Taylor series. Accuracy of this method depends on the accuracy of estimation of the exponential function with finite order of the Taylor series. For small-amplitude waves and for the narrow wave spectrum, such accuracy is evidently high. Efficiency of the HOSM approach for simulation of the low-frequency, energy-containing part of the spectrum should be good. After including realistic input and dissipation terms, the HOSM can be used successfully for research purposes for wave simulation in suitable domains. However, for a case of the broad wave spectrum, which contains many wave modes, the order of the Taylor series should increase. It is likely that the HOSM model should have limitations when simulating evolution of waves with broadband spectrum for large domains.

Direct methods for 3D waves include the elliptic boundary layer problem solved by finite-difference methods. Such approaches to simulation of the unsteady free surface flows based on full equations have been under development for at least three decades (see, e.g., Bingham and Zhang 2007). The main advantage of this method is that it is based on initial equations being transformed into the surface-following coordinate system. All of the papers of this group were mostly dedicated to technical applications of the water wave theory, for example, to calculations of dynamic load on submerged bodies or to simulation of the wave dynamics in a domain with complicated shape. A long-term evolution of such water motions was never simulated; this is why exact conservation of energy was not the main priority of such models. Applicability of these models for investigations of nonlinear properties of sea waves is also uncertain.

Another method for solution of the 3D equations is based on presentation of velocity potential as a sum of linear and nonlinear components (Chalikov et al. 2014). Advantage of this presentation is due to the nonlinear components having simple vertical structure and being about two decimal orders smaller than linear components. This model is most suited for simulating long-term evolution of multi-mode wave field.

All phase-resolving models can be extended by including algorithms for input and dissipation of energy. This makes the whole approach highly efficient for investigation of physics of realistic surface ocean waves.

3. Spectral models

Evolution of wind-generated waves in water of finite depth can be described by the wave action $N = F/\omega$ balance equation

$$\frac{\partial N}{\partial t} + \nabla \cdot [(\mathbf{c}_g + \mathbf{U})N] + \nabla_{\mathbf{k}} \cdot [\mathbf{c}_{\mathbf{k}} N] = \frac{I + L + D + B}{\omega}. \tag{1}$$

where $F(\omega, \mathbf{k})$ is wave energy density spectrum, ω is intrinsic (from the frame of reference relative to any local current) radian frequency, $\mathbf{k}$ is wavenumber (bold symbols signify vector properties). In linear case, temporal and spatial scales of the waves are linked through the dispersion relationship

$$\omega^2 = gk \cdot \tanh(kd) \tag{2}$$

where g is the gravitational constant and d is water depth. Note two asymptotic versions of this equation, which signify the deep water when $kd \gg 1$, i.e., when Equation (2) becomes a simple algebraic function $\omega^2 = gk$, and the shallow water when $kd \ll 1$, i.e., $\omega/k = \sqrt{gd}$. The latter indicates nondispersive environment when phase speed of the waves $c = \omega/k$ does not depend on the wavelength and frequency.

The left-hand side of Eq. (1) represents time–space evolution of the wave action density as a result of the energy source terms on the right. On the left, $\mathbf{c}_g$ is group velocity, $\mathbf{c}_{\mathbf{k}}$ means the spectral advection velocity, and $\mathbf{U}$ is the current speed. ∇ here is the horizontal divergence operator, and $\nabla_{\mathbf{k}}$ is such operator in spectral space.

On the right, source terms are phenomenologically represented by atmospheric energy input from the wind, I; nonlinear interactions of various orders within the wave spectrum, L, whose role is to redistribute the energy within the spectrum; dissipation energy sinks, D; wave-bottom interaction processes, B; and more sources are possible in specific circumstances. Note that all the source terms, as well as the group and advection velocities, and the advection current are spectra themselves. We refer the reader to Cavaleri et al. (2007) for further details.

In the context of wave–ocean interactions, both left-hand side and right-hand side have their meaning. $\mathbf{U}$ is the advection current velocity that in general should depend on the vertical profile of the ocean current and on steepness (nonlinearity) of the waves at different spectral scales, i.e., to describe coupled wave-current nonlinear exchanges (but in practice is usually approximated by the speed of drift current). $\mathbf{c}_g$ defines horizontal propagation of wave energy; $\mathbf{c_k}$ means the spectral advection velocity to represent refraction (turning of wave crests) and change of wavelength when waves enter finite depths (shoaling), or the dispersion relation (2) is modified due to some other circumstances such as, for example, presence of the surface ice.

On the right, L is a conservative term, i.e., its integral is zero, but the other integrals define energy fluxes in and out the wave system.

$$E_I = \int I(\omega, \mathbf{k}) d\omega d\mathbf{k} \tag{3}$$

is the total flux of energy from the wind to the waves. Note that, depending on the relative speed of wind $\mathbf{U}_{10}$ and wave speeds $\mathbf{c}(\omega, \mathbf{k}) = \omega/\mathbf{k}$, contributions to the total flux can be both positive (from the wind to the waves if $U_{10} > c$) and negative (from the waves to the wind if $U_{10} < c$). In the tropics, for example, where the wave climate is dominated by swells produced at high latitudes, the local winds are typically light and therefore the wind climate can be actually dominated by wave-induced winds (Hanley et al. 2010).

It should be noted that the energy input to the waves is generally accepted as a purely atmospheric exchange. In principle, however, energy input from the ocean side to the surface waves of scales accommodated in Eq. (3) is perceivable. For example, upper-ocean currents, tides, or internal waves can provide such dynamics. Given the amount of energy stored in the ocean movements, this could have large impacts on surface wave fields, even if localized, but it would be fair to say that it has not been considered by the wave-ocean modeling community in practical terms.

Integrating the momentum-input spectrum gives the total momentum flux

$$\boldsymbol{\tau}_w = \int I(\omega, \mathbf{k})/\mathbf{c}(\omega, \mathbf{k}) \mathrm{d}\omega \mathrm{d}\mathbf{k} = \int I(\omega, \mathbf{k}) \frac{\mathbf{k}}{\omega} \mathrm{d}\omega \mathrm{d}\mathbf{k} \tag{4}$$

which is an important measure of wind-wave interactions (see, e.g., Tsagareli et al. 2010). Together with the tangential viscous stress τ_ν, it forms the total wind stress at the ocean surface

$$\boldsymbol{\tau} = \boldsymbol{\tau}_w + \boldsymbol{\tau}_\nu, \tag{5}$$

and this stress is known independently (usually through empirical parameterizations of the so-called drag coefficient) and thus can be used as a constraint or for validation of the wind input term I. On the other hand, the total stress is often the main, if not the only, property that expresses dynamic exchanges in large-scale air-sea models. Apart from situations of light winds, the wave-induced form drag (4) provides a dominant contribution to this total stress (e.g., Kudryavtsev et al. 2001) and, thus, if the wave-model physics is well defined and validated, such models can provide explicit rather than empirical estimates of fluxes for General Circulation Models if those are appropriately coupled with wave models.

The dissipation function D has a similar meaning in the context of wave-ocean dynamic exchanges, but with some essential distinctions. First, the integral

$$E_D = \int D(\omega, \mathbf{k})\mathrm{d}\omega\mathrm{d}\mathbf{k} \tag{6}$$

is the total flux of energy out of the wave field. The energy passed to the ocean is largely spent on generating turbulence near the surface and on work against buoyancy forces acting on bubbles injected in the course of the wave breaking.

Unlike the input, however, which only occurs on the air side of the interface, the loss (6) can go both to the ocean below and to the atmosphere above the ocean surface. Numerical simulations of Iafrati et al. (2013), for example, showed that up to 80% of wave energy due to breaking can be actually dissipated through the atmospheric turbulence.

The momentum-loss integral of dissipation function gives the so-called radiation stress,

$$\boldsymbol{\tau}_r = \int D(\omega, \mathbf{k})\frac{\mathbf{k}}{\omega}\mathrm{d}\omega\mathrm{d}\mathbf{k} \tag{7}$$

which is presumed to be going to the currents (although some of it may in fact be going back to the wind or to the bottom in shallow areas). It should be pointed out that, in the present wave models, radiation stress is parameterized in terms of wave-height difference along the propagation direction. Obviously, such parameterization does describe the energy dissipation, and can then be used to estimate the momentum loss, but only in the areas where dissipation (6) is much larger than the energy input (3), i.e., usually in shallow waters. In deep water, the mean wave height is not a proxy for the energy loss. It may in fact grow under wind action, or not change if this action is balanced by the white-capping dissipation, but the integral (7) and hence the radiation stress, is not zero.

Wave-ocean-bottom interactions in finite depths, depicted by term B in (1), are very rich. Finite depths are characterized by the condition of $kd \sim 1$ (wavelength is comparable to the water depth d) and, their further subdivision, shallow nondispersive environment waters by $kd \ll 1$ as mentioned above. Dispersive-wave nonlinear dynamics slow down in finite and shallow depths, weaken, or cease, but other nonlinear behaviors come into existence.

Wave exchanges with the bottom include bottom friction, formation of ripples, sediment suspension, and transport if the sea bed is sandy, generation of bottom waves if the bottom is muddy, percolation, among others. Long-shore and cross-shore currents, rip currents result from radiation stresses (7), infragravity waves are produced by combined action of wave breaking and nonlinear wave groups, which can be subsequently reflected back to the deep ocean or trapped by coastal bays.

a. Wind input

It is generally accepted that the energy flux from the wind to the waves is due to work of pressure stresses correlated with the surface slope (e.g., Young 1999):

$$I = \frac{1}{\rho_w g} \overline{p \frac{\partial \varsigma}{\partial t}} \tag{8}$$

where ρ_w is water density, p is atmospheric pressure at the surface, ζ is surface elevation, and overbar represents an average with respect to time. Pressure has to have a component in quadrature with surface elevations, in order for this mechanism to work—note that the flux can be both positive and negative as mentioned above. There are a number of theories that try and explain the nature of such wave-induced pressure, ranging from the phenomenological (Jeffreys 1924, 1925) through the quasi-laminar (Miles 1957, 1958) to the fully turbulent (Belcher and Hunt 1998).

As such, this is the only source function in (1) which can be directly measured. The omnidirectional quadrature spectrum provides fractional growth rates $\gamma(\omega)$ so that

$$I(\omega) = \gamma(\omega)\omega F(\omega). \tag{9}$$

The measurements, however, are challenging. They need to be taken in a wave-following system, in order to keep the air-pressure probe dry and still take readings near the surface below wave crests. Up to date, only a handful of such experiments, field and laboratory, have been done and only one field experiment carried out such measurements in a broad range of conditions including strong winds (Donelan et al. 2005, 2006). Even then, only frequency spectra have been measured and directional distribution of the wind input remains unknown, although is imposed in models, usually as a cosine distribution following the suggestion of Plant (1982) (see, however, Shabani et al. 2016).

It should be mentioned that the input term (8) can only be applied to the water surface once the waves already exist. Practical importance of the very early stages of wave development is small, but for spectral models of wave forecast, which often need a cold start from the flat surface, this is of principal significance—if there is no initial spectrum in (9), the wind input will always remain zero. The generally accepted basic theory for this kind of wave generation is that of Phillips (1957), which relies on turbulent vortexes on the air side of interface, propagated by the wind at a phase speed of waves being generated. This theory has never been verified experimentally, and other mechanisms of the initial wave generation such as, for example, Kelvin-Helmholtz instability, are also possible.

As mentioned above, for surface waves in the range of wavelengths from centimeters to kilometers, the wind (directly or indirectly) is regarded as the only source of energy and momentum, but it is not impossible that at some circumstances the waves can gain energy from other interactions.

b. Dissipation sinks

Dissipation function (or rather functions) is as important for modeling the ocean waves, as the input term. As soon as the waves appear, the dissipation mechanisms activate, and the wave growth and evolution is always a balance between energy inputs and outputs. Here we refer to a review book on this topic (Babanin 2011).

Unlike the only energy input, dissipation is recognized as a variety of superposed processes, not necessarily related, which effectively require separate dissipation terms in (6), i.e., $D = D_1 + D_2 + D_3 \ldots$ The primary dissipation of wind-forced waves is due to wave breaking, which is an intermittent, rapid, and highly nonlinear process. It happens at the scale of tens of wave periods, but an individual breaking wave loses its energy and momentum, slowly accumulated due to the wind forcing, within a fraction of wave period. This dissipation is only active when mean wave steepness (or spectral density) is above a certain limit $F_{th}(\omega)$—low steepness waves do not break.

The wave-breaking dissipation is routinely called "white-capping," but it should be remembered that micro-breakers do not produce whitecaps, and they are actually much more widespread (e.g., Jessup et al. 1997). Another feature of the white-capping dissipation that distinguishes it from the wind input is that it is not local in wavenumber–frequency space: waves at long scales up to peak frequency ω_p affect the breaking and dissipation at short scales, and hence the dissipation term has a cumulative integral of the wave spectrum and is very different from the input term (9):

$$D(\omega) = -\gamma_1(\omega)\omega(F(\omega) - F_{th}(\omega)) - \int_{\omega_p}^{\omega} \gamma_2(\omega)\omega(F(\omega) - F_{th}(\omega))d\omega. \quad (10)$$

Eq. (10) is an example that accommodates both the threshold and cumulative behavior, and different implementations of this term now exist including its dimensionless and nonlinear forms (e.g., Ardhuin et al. 2010; Rogers et al. 2012).

Once the wind forcing is reduced or ceased, or the waves propagate outside the storm area, they become swell and the breaking stops. Therefore, swell dissipation should be of completely different nature. Unlike the white-capping dissipation, it does not have threshold and is not intermittent. It is a continuous and slow process; swells can cross entire oceans subject to gradual decay due to such interactions. The main mechanisms usually considered as such dissipation are generation of turbulence by surface waves, either in the water (Babanin 2006, 2011) or on the air side (Ardhuin et al. 2009), while the viscous attenuation of water waves is regarded negligible. It should be mentioned that the so-called swell dissipation is present in the wind-generated waves also, albeit relatively small by comparison with white-capping, but can become dominant, for example, at the peak of Pierson-Moscowitz spectrum when the waves are not steep enough to break.

While wave-breaking and swell dissipations are ever-present, other wave-attenuation mechanisms can be very important locally and in specific circumstances. Among such mechanisms, for example, is interaction of waves with adverse or oblique winds (e.g., Donelan 1999). Sometimes, it is treated as negative wind input, but essentially it causes loss of energy from the surface waves and hence their dissipation.

c. Nonlinear interactions and other source terms

In the context of wave modeling, the term nonlinear interactions is loosely (but not always strictly) attributed to the energy exchanges within wave system (i.e., wave fields or wave spectrum, depending on whether physical space or Fourier space is being modeled), which do not bring energy in or out of the system, but rather redistribute it within. For example, the wind input (9) can be nonlinear if, for example, the growth rate $\gamma(\omega)$ is a function of spectrum $F(\omega)$, such input function was proposed by Donelan et al. (2006), but it would not be held responsible for nonlinear interactions in the traditional context of spectral wave modeling. Neither would be a dissipation function, even though in some spectral models the redistribution of wave energy within the wave spectrum is explicitly attributed to wave breaking (e.g., Donelan et al. 2012).

The most traditional perception of the nonlinear-interaction term in spectral wave models is that due to resonant four-wave interactions, as proposed originally by Hasselmann (1962), but these days usually derived from the Zakharov (1968) equation. Unlike the input and dissipation source functions, which are highly empirical, this term is described analytically and usually perceived as conservative and known exactly. In reality, however, the exact integral equation that describes this term is computationally very expensive and therefore, even with the modern-day powerful computing facilities, cannot be employed for operational forecast. Instead, variations of the Discrete Interaction Approximation method (DIA; Hasselmann et al. 1985) are routinely used for practical needs. Moreover, only a limited number of frequencies around the spectral peak are usually used in such models, and hence some nonlinear fluxes that are supposed to transfer energy to small scales with the spectrum are lost, and therefore practical implementations of the nonlinear term are not strictly conservative either.

Quasi-resonant nonlinear interactions are not depicted by the Hasselmann equation, but they are by the Zakharov Equation, and respective source terms suitable for spectral models have also been derived (Annenkov and Shrira 2006; Gramstad and Stiassnie 2013) and implemented in spectral models (Gramstad and Babanin 2016). Not only do they account for some missing part of energy exchanges, but most essentially for missing dynamics, the so-called modulational instability (Zakharov 1966 1967; Benjamin and Feir 1967; Yuen and Lake 1982). This instability is responsible for large individual waves and wave breaking, and is impaired in directional wave fields (e.g., Onorato et al. 2001; Janssen 2003). It is of great significance, however, if the angular spread is narrow enough, which can be the case, for example, for waves on currents (e.g., Shrira and Slunyaev 2014).

The resonant and quasi-resonant interactions cease as the water shallows and the waves become less dispersive. Other nonlinear dynamics, however, emerge and gain significance, such as triad interactions (e.g., Young and Eldeberky 1998) or infragravity waves (e.g. Herbers et al. 1995). The latter, however, takes us far away from the energy-conserving nonlinear exchanges and into a broad suit of new dynamics important in finite depths. Here we will refer to Komen et al. (1994), Cavaleri et al. (2007), Holthuijsen (2007) for reviews of these specialized topics. Another broad range of new dynamics that gain attention lately is wave-ice interactions (see e.g., Montiel et al. 2016) for a recent update. As already mentioned, many more energy source and sink terms are possible and used in specific circumstances.

4. Waves on currents

It would be fair to say that the wave-current interactions are the least well-described physical mechanisms in the models for wave forecast. While linear wave effects are usually included, nonlinear effects are usually not, and some of them are not even known or well understood. We refer to Mei (1983) and a recent review by van der Westhuijsen (2016) of the wave-current problem.

The linear effects contain changes to the frequency, wavenumber and direction, including nonstationary and nonuniform currents, Doppler shifting and refraction, which are accommodated in the action balance equation (1). For practical applications, changes of the relative wave speed with respect to the wind is of significance (Haus 2007), and are also incorporated in recent versions of wave forecast and hindcast models.

In equation (1), however, with source and sink functions such as input (9) and dissipation (10), the linear kinematic effects due to currents should cause significant (and nonlinear) dynamic consequences. The wind input, for example, is influenced both by variation of steepness (as a result of wavenumber change) and refraction. The steepness affects growth rates $\gamma(\omega)$ (Donelan et al. 2006), and the direction with respect to the wind changes the relative wave speed (Haus 2007). As a result, the input becomes a nonlinear function of the wave spectrum $F(\omega)$.

Most essentially, however, the wave steepness controls the wave breaking and hence the dissipation function (10). This is a very complex behavior, nonlinear, thresholded, and cumulative, as described above. Besides, wave breaking further impacts on the atmospheric boundary layer (Iafrati et al. 2013) and the wind input (Babanin et al. 2007). Thus, consequences of waves entering, propagating along or across or leaving the currents, even in the context of purely linear kinematics of such waves, are quite complicated. Since, moreover, practical implementations of source and sink terms (9) and (10) are highly empirical and usually obtained and tested for some simple conditions of wave evolution, their validity when modeling waves on the currents is still largely an open question.

One specific condition of waves on currents, responsible for dangerous seas and even (supposedly) rogue waves, is when waves are entering opposing accelerating currents. Such situations are regular occurrence, for example, at the southern tip of Africa where strong Agulhas Current down the east coast meets also strong swells propagating north from the Southern Ocean. Technically speaking, any tidal inlet can be subject to such situation daily, depending on wind (and hence wave) direction during the ebb tide.

Waves propagating into an opposing current gradient with increasing strength will experience an increase in the intrinsic frequency ω (reduction in wavelength according to (2)), which causes an increase in wave height due to (1). Hence, wave steepness is rapidly growing, as well as wave breaking. This effect is known and well appreciated, but is not well understood and described analytically, and performance of wave models in such conditions is usually poor due to inadequate dissipation functions (van der Westhyijsen 2012). It should be noted that a following current that decelerates also causes wave steepening.

Wave steepening also brings about various nonlinear dynamics not accommodated in phase average models. In particular, modulational instability, which can produce high individual waves, even rogue waves, depends strongly on steepness (e.g., Tulin and Waseda 1999) and, furthermore, becomes very rapid in presence of currents (e.g., Chawla and Kirby 2002).

An asymptotic situation of waves meeting an opposing current is when velocity of the current approaches the wave group velocity, and thus, according to the linear theory, the waves should be blocked. In reality, modulational instability or other nonlinear effects can cause downshifting of the wave energy, and hence penetration of waves beyond the blocking point (Babanin et al. 2011).

The ocean surface is two-dimensional, and currents also have lateral velocity gradients. Respective refraction of currents can express itself in wave focusing and trapping, which is a phase-resolving effect and would not be revealed in outcomes of modeling the action balance equation (1) (see, e.g., Lavrenov (1998) or Zhang et al. (2009) more recently). Nonlinear dynamics of wave-current trapping can also be of importance (Shrira and Slunyaev 2014).

In wave-current research, currents are often considered depth-integrated or uniform vertically. Such approximation works well for tidal currents or in shallow water, but in the open ocean surface currents are also ubiquitous. Such currents will have an exponential vertical velocity gradient, which is usually different from the vertical gradient of the wave orbital velocity, and therefore nonlinear wave-current interactions become inevitable.

5. Effects of waves on currents

Surface waves are essentially a 3D phenomenon. The vertical exponential profiles for the orbital velocity and mean Stokes drift are a necessary part of the propagating surface oscillations, extending down into the water at the scale of wavelength ($\sim$200 m in the open ocean). If, for example, a plunging wavemaker in a wave tank produces water oscillations with a linear vertical velocity profile, the wave train will not propagate forward like this and will adjust it to an exponential vertical profile to fit the corresponding wavenumber:

$$v = v_0 \exp(-kz) = a_0 \omega \exp(-kz) \tag{11}$$

where a_0 and v_0 are the surface amplitude and orbital velocity, respectively (in linear sense).

Any modifications to the surfaces waves imply respective changes to the velocity profile, and the reverse is also true. If the orbital velocity profile is modified, due to presence of the currents, for instance, this should reflect on the surface wave fields.

Energy and momentum of the wave motion by definition is concentrated near the surface at the wavelength scale, and therefore are small by comparison with energy of, for example, tides or geostrophic currents. As a result, even weak nonlinear exchanges between surface waves and such currents can impact significantly on wave fields. For surface currents, however, and for currents in shallow waters comparable to wave lengths, the wave-current dynamic exchanges can influence the mean flow or ocean circulation locally.

a. Depth-integrated effects

In the depth-integrated approach, the continuity and momentum equations (the latter are for horizontal and vertical velocities separately, with account for viscous stresses, but not turbulent stresses), are written for the depth-integrated mean current velocity U. The mean current is averaged over wave period, and instantaneous current u includes superposed wave oscillation $\tilde{u}$:

$$u_i(\mathbf{x}, \mathrm{z}, \mathrm{t}) = \mathrm{U}_i(\mathbf{x}, \mathrm{t}) + \tilde{\mathrm{u}}_i(\mathbf{x}, \mathrm{z}, \mathrm{t}). \tag{12}$$

Here, index i signifies coordinate axes, $\mathbf{x}$ is horizontal vector, z vertical coordinate, t is time. z-integration is from the bottom to the free surface (surface elevation ζ). If averaged over the wave period, the last term is zero for linear waves, but for nonlinear waves the Stokes drift becomes a part of the mean flow. We refer to Mei (1983), van der Westhuijsen (2016), and will mention here that kinematic and dynamic boundary conditions are applied at the free surface and the bottom. The surface stress τ^s is composed of the atmospheric stress (including wind stress) and surface tension, applied to the wavy surface, and the bottom stress τ^b is defined by the no-slip condition (bottom can be sloped).

The depth-integrated continuity equation develops into:

$$\frac{\partial < \varsigma >}{\partial t} + \frac{\partial}{\partial t}[(d + < \varsigma >)U_i] = 0 \tag{13}$$

where $< \ldots >$ denotes time averaging over several wave periods. The wave-averaged form of the horizontal momentum equation (if viscous stress, surface, and bottom slopes are all assumed small) becomes

$$\rho_w(d + \bar{\varsigma})\left[\frac{\partial U_i}{\partial t} + U_j \frac{\partial U_i}{\partial x_j} + g\frac{\partial \bar{\varsigma}}{\partial x_j}\right] + \frac{\partial S_{ij}}{\partial x_j} = \tau_i^s - \tau_i^b. \tag{14}$$

Here, i and j are horizontal coordinates, $\bar{\varsigma}$ is the mean free surface. S_{ij} is the radiation stress tensor, which, if the viscosity is neglected, is written out as follows:

$$S_{ij} =< \int_{-d}^{\varsigma} (\rho_w \tilde{u}_i \tilde{u}_j + p\delta_{ij}) dz > -\frac{\rho_w g}{2}(h + \bar{\varsigma})^2 \delta_{ij}. \tag{15}$$

In (15), δ_{ij} is the Kronecker symbol, the first term in the integrant represents the amount of momentum transferred by the water particles in the wave orbits, the second the transfer due to normal pressure, and the last term represents the hydrostatic pressure. Finally, within the same assumptions, the vertical velocity

$$U_z = 0 \tag{16}$$

(no mean vertical flow).

Note that the hydrodynamic model above, and therefore the radiation stress formulation Eq. (15), do not explicitly take wave breaking into account. Instead the momentum lost by breaking waves is exchanged between the wave field and the mean flow (see also Garrett 1976; Smith 2006).

Confusions exist in the oceanographic literature in this regard because, indirectly, the wave breaking may be accounted for through the variation of the mean water level. These are conditions when the breaking is strong and frequent (scale of tens of wave periods or less), and therefore the wave energy (and height) cannot be compensated by the wind input (scale of thousands of wave periods). These are typically situations of rapid shoaling.

Such a situation was first discussed by Longuet-Higgins and Stewart (1954) and obtained numerous applications in coastal modeling and engineering. They considered a one-dimensional case (x-axis) of a monochromatic wave train approaching a mild-sloped beach, on which the waves shoal and ultimately break. Here, the wave breaking is not explicitly included, but is responsible for variation of the wave height $\bar{\varsigma}$. As a result, the non-uniform cross-shore distribution of the radiation stress is formed

$$S_{xx} =< \int_{-d}^{\varsigma} (\rho_w \tilde{u}_x^2 + p) dz > -\frac{\rho_w g}{2}(d + \bar{\varsigma})^2. \tag{17}$$

If stationary condition is reached (mean horizontal flow $U_x = 0$), wind and bottom stresses are neglected, Eq. (14) becomes

$$\frac{dS_{xx}}{dx} + \rho_w g(d + \bar{\varsigma})\frac{d\bar{\varsigma}}{dx} = 0. \tag{18}$$

This gives us the direct relationship between the gradient of the cross-shore radiation stress dS_{xx}/dx and the slope of the mean free surface $d\bar{\varsigma}/dx$ Hence, for a given wave field, by integrating (18) from the offshore towards the coast, a cross-shore gradient in the water level can be obtained. It is known as wave-induced set down and set-up.

Van der Weshuijsen (2016) distinguishes two zones. The first is leading from the offshore up to the breaking point where energy gains or losses are typically gradual (e.g., wind input or bed friction), and the second from the breaker line to the shore, where the waves rapidly lose their energy (and height) due to breaking.

In the first zone, the mean surface $\bar{\varsigma}$ becomes negative (lowering of the mean surface level), with increasingly negative values as the total depth reduces. This phenomenon is referred to as wave-induced set-down, and occurs because the radiation stress S_{xx} increases with decreasing depth, in this non-dissipative approach. Considering the region from the breaking point shoreward, as long as the bed slope stays positive, slope of the mean free surface will also be positive, a phenomenon known as wave-induced setup.

The other well-known application of non-dissipation wave radiation stresses in shallow waters is the phenomenon of longshore current generation, also introduced by Longuet-Higgins (1970a, b). This process is two-dimensional, with y-axis oriented alongshore. The incoming wave field is uniform along the length of the beach (y-axis), and the waves refract towards shore normal and ultimately break as the depth reduces.

Considering that the flow and water level will be uniform along the infinitely long coast ($\partial\bar{\varsigma}/\partial y = 0$), assuming that the cross-shore flow will be zero ($U_x = 0$) and neglecting the surface stress τ^s (but not the bottom stress τ^b in this exercise), we have for the longshore current U_y the following balance

$$\frac{\partial}{\partial t}[\rho_w(d+\bar{\varsigma})U_y] = -\frac{\partial S_{yx}}{\partial x} - < \tau_i^b > . \tag{19}$$

Therefore, as waves obliquely approach a beach from deep water, the negative cross-shore gradient of the radiation stress term $\partial S_{yx}/dx$ (mainly due to wave breaking) will tend to accelerate the longshore current U_y.

It is now clear why τ^b was not neglected initially like in the case of wave setup above. The acceleration of U_y will in turn increase the bed stress term τ^b:

$$\tau^b = \frac{1}{2} f_w \rho_w U^2 \tag{20}$$

where f_w is the friction factor. This will continue until a steady longshore current develops, at which time the left-hand side of (19) goes to zero, and the two terms on the right-hand side are in balance.

b. Depth-varying description

In the case of depth-integrated current above, the Eulerian approach was applied and the mean current and wave motion were split. This is not easy to do if the mean current is depth varying. Typically, description of waves in such currents is accomplished by using the Generalized Lagrangian Mean (GLM) method (Andrews and McIntyre 1978; Jenkins 1989; Groeneweg and Klopman 1998; Mellor 2003). For applications in numerical models, Ardhuin et al. (2008), Bennis et al. (2011) modified GLM as a quasi-Eulerian approach.

Conceptually and mathematically GLM is a hybrid of Eulerian and Lagrangian approaches (see the original papers and review by van der Westhuysen (2016)). While in an Eulerian description the control volume is defined as having a fixed position, in the GLM approach the control volume is displaced by the disturbance motion. So it is neither purely Eulerian nor Lagrangian system.

The generalized Lagrangian mean is defined as the mean at its disturbed position $\mathbf{x} + \xi(\mathbf{x}, t)$:

$$\bar{\phi}^L(\mathbf{x}, t) = < \phi(\mathbf{x} + \xi(\mathbf{x}, t), t) >, \tag{21}$$

i.e., following a single fluid parcel, in contrast to the Eulerian mean that is evaluated at its original position. The Stokes correction is defined as the difference between the Lagrangian and Eulerian means. For infinitesimal waves, the disturbance displacement is assumed to obey

$$< \xi(\mathbf{x}, t) > = 0 \tag{22}$$

and the Lagrangian disturbance velocity is defined as

$$\mathbf{u}^l(\mathbf{x}, t) = \mathbf{u}(\mathbf{x} + \xi(\mathbf{x}, t), t) - < \mathbf{u}(\mathbf{x}, t) >^L. \tag{23}$$

Finally, the temporal derivative of the displacement ξ is defined as the rate of change following the mean flow, so that

$$\mathbf{u}^l = \left(\frac{\partial}{\partial t} + < \mathbf{u} >^L \cdot \nabla \right) \xi. \tag{24}$$

These equations form the basis of the GLM theory and are then used to derive Langrangian mean expressions for the continuity of mass and momentum.

6. Numerical examples

Gravois et al. (2012) and Allard et al. (2014) describe a validation of the Navy's coupled ocean-wave modeling system, COAMPS (Hodur 1997), including a test case for southeast Florida, where shore-based high frequency (HF) radars are used as ground truth for evaluation of the wave model output. They find that by including the effects of surface currents in the modeling, there is better correlation with the observations. This is, of course, to be expected. One less obvious—and arguably more interesting—conclusion is that for cases in which swells must pass through narrow apertures, or at angles that are very oblique to the shoreline, a circumstance that is not uncommon for that area—the question of whether or not swells arrive at nearshore locations is strongly dependent on the existence of currents, meaning that if currents are omitted, swells that do not arrive at all may be predicted (false positive), or swells that do in fact arrive may be missing in the model (false negative). This is illustrated in Figures 1–3.

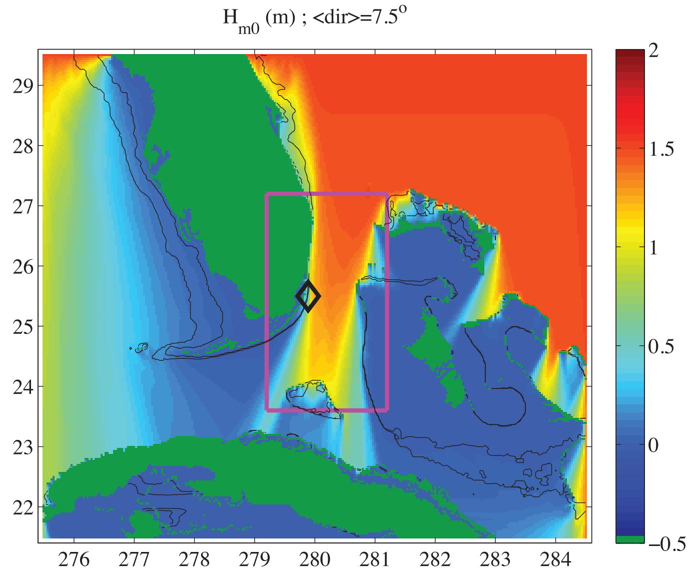

Figure 1. Idealized propagation in wave model with all incoming wave energy coming from boundaries with propagation direction of theta = 7.5° (from north to northeast). The magenta rectangle indicates a nearshore wave model nest (Fig. 2). Black lines indicate 10 m and 20 m isobaths. Black diamond indicates location of *in situ* data used for validation by Gravois et al. (2012).

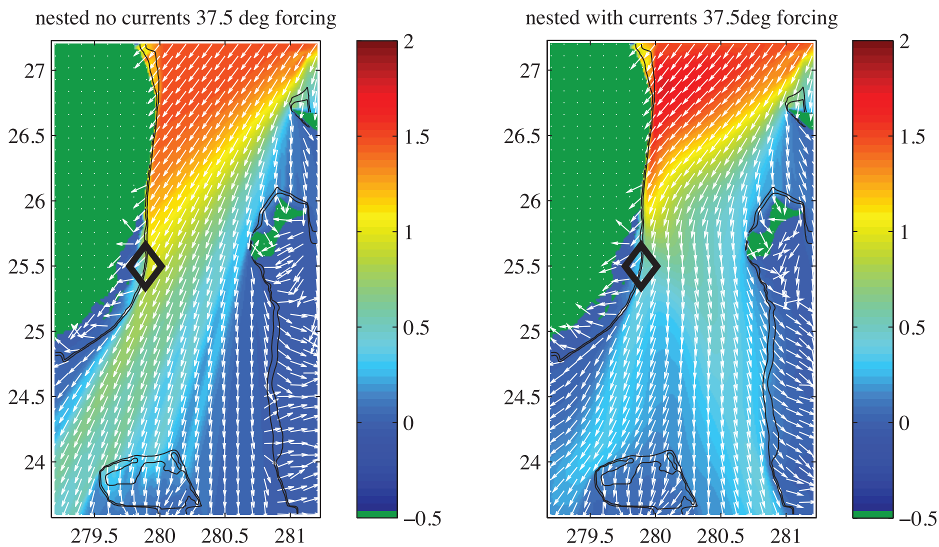

Figure 2. Idealized propagation in a nearshore wave model nest, in which all incoming wave energy has propagation direction theta = 37.5° (from northeast). Black diamond and black contours: see explanation in the caption of Figure 1. Arrows indicate mean wave direction. Left panel: without currents included in forcing. Right panel: with currents included in forcing.

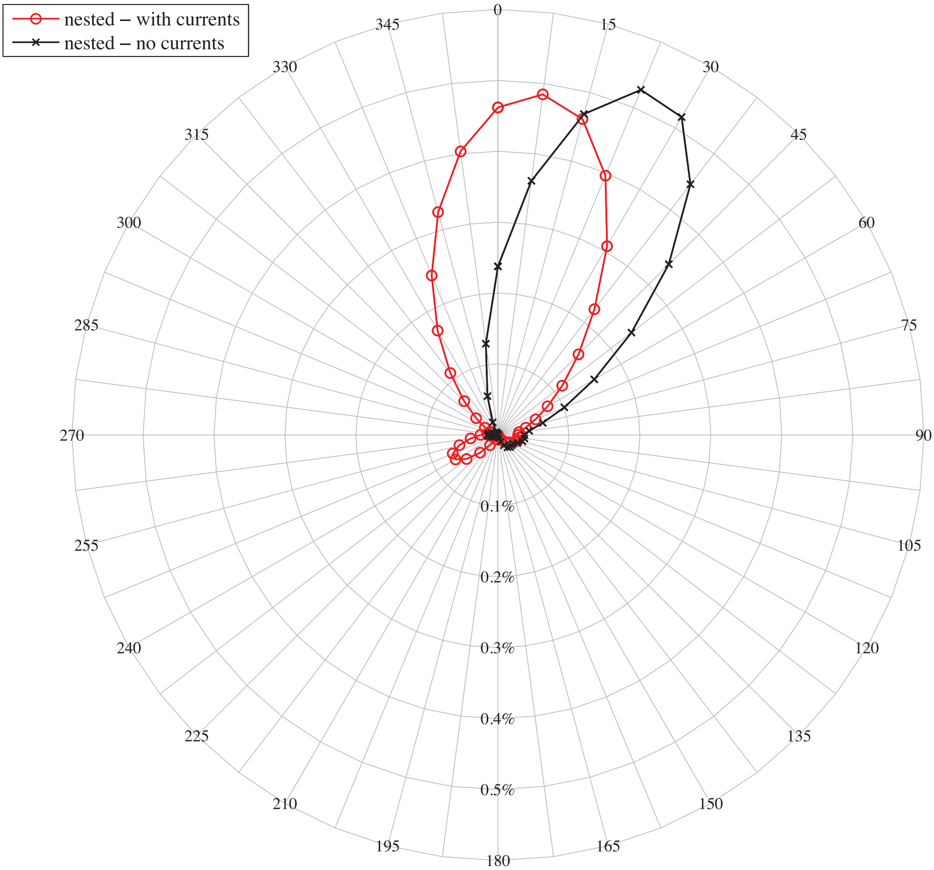

Figure 3. Polar plot of transmission ratio as predicted by a wave model: [wave height at location of *in situ* measurements (location marked in Figures 1 and 2)] / [wave height at offshore boundary of wave model]. Black denotes case with no currents and red denotes case with currents included in wave model forcing.

7. Discussion and summary

Wave-current interactions are a topic of great challenge in fluid mechanics and oceanography, and of important practical significance. Its advanced modeling, however, is yet to come.

Practical models, used for wave forecast, only accommodate linear effects due to waves propagating on or across the currents. Even then, the linear changes to wave kinematics cause nonlinear dynamic consequences in wave growth and dissipation that are not well described and understood, and performance of respective models in this context is poor. Nonlinear behaviors of waves on currents are usually not accounted for or even unknown.

Waves, in turn, can substantially influence the surface currents through their Stokes drift, and even more so through radiation stress (7) due to the momentum lost in wave breaking. These influences are largely missing in modeling the ocean circulation of open oceans, and they certainly need to be reinstated because of their importance, for example, in search and rescue missions. In principle, this can be done by coupling the wave models with the circulation models, but this should be done with caution. The present-day wave forecast models were not designed to produce correct fluxes for input (4) and dissipation (7), but rather an approximate balance of those in order to predict the resulting wave growth and wave height reasonably well.

Most essential is the role of waves in influencing and even producing currents in finite depths. Here wave breaking is extensive, and this role is known and well accounted for in coastal circulation and coastal engineering models.

Apart from direct wave-current applications, a more complex suit of wave-current-turbulence interactions, both on the atmospheric and ocean sides, is gaining attention over the recent years. While energies involved are small in the context of changes to the waves or currents, they may play an additional and missing feedback in the atmospheric boundary layer, where airflow then generates both surface waves and currents, and in the upper ocean mixing.

Overall, wave-current interactions are the last loose element of physics of wave forecast models, and, to an extent, of ocean circulation modeling. A rich variety of dynamics are involved, some of which are not well understood, and complicated mathematics to describe such dynamics, make it an exciting research field at the boarder of wave, ocean and meteorological applications.

Acknowledgments. A.V. Babanin acknowledges support of the Australian Research Council through grant DP170100851. WER was funded by the Office of Naval Research under Program Element 0602435N.

REFERENCES

Allard, R., E. Rogers, P. Martin, T. Jensen, P. Chu, T. Campbell, J. Dykes, et al. 2014. The US navy coupled ocean-wave prediction system. Oceanography, *27*, 92–103. doi: 10.5670/oceanog.2014.71

Andrews, D. G., and M. E. McIntyre. 1978. An exact theory of nonlinear waves on a Lagrangian-mean flow. J. Fluid Mech., *89*, 609–646. doi: 10.1017/S0022112078002773

Annenkov, S. Y., and V. I. Shrira. 2006. Role of non-resonant interactions in the evolution of nonlinear random water wave fields. J. Fluid Mech., *561*, 181–207. doi: 10.1017/S0022112006000632

Ardhuin, F., N. Rascle, and K. A. Belibassakis. 2008. Explicit wave-averaged primitive equations using a generalized Lagrangian mean. Ocean Model., *20*, 35–60. doi: 10.1016/j.ocemod.2007.07.001

Ardhuin, F., E. Rogers, A. Babanin, J.-F. Filipot, R. Magne, A. Roland, A. van der Westhuysen, et al. 2010. Semi-empirical dissipation source functions for ocean waves. Part I: Definitions, calibration and validations. J. Phys. Oceanogr., *40*, 1917–1941. doi: 10.1175/2010JPO4324.1

Ardhuin, F., B. Shapron, and F. Collard. 2009. Observation of swell dissipation across oceans. Geophys. Res. Lett., *36*, L06607. doi: 10.1029/2008GL037030

Babanin, A. 2011. Breaking and Dissipation of Ocean Surface Waves. Cambridge: Cambridge University Press, 480 p.

Babanin, A. V. 2006. On a wave-induced turbulence and a wave-mixed upper ocean layer. Geophys. Res. Lett., *33*, L20605. doi: 10.1029/2006GL027308

Babanin, A. V., M. L. Banner, I. R. Young, and M. A. Donelan. 2007. Wave-follower field measurements of the wind-input spectral function. Part III: Parameterization of the wind-input enhancement due to wave breaking. J. Phys. Oceanogr., *37*, 2764–2775. doi: 10.1175/2007JPO3757.1

Babanin, A. V., H.-H. Hwung, I. Shugan, A. Roland, A. van der Westhuysen, A. Chawla, and C. Gautier. 2011. Nonlinear waves on collinear currents with horizontal velocity gradient, *in* Proceedings of the 12th International Workshop on Wave Hindcasting and Forecasting and Third Coastal Hazard Symposium, Kona, Hawaii, October 30–4 November 2011, JCOMM-TR-067, JCOMM Technical Report Series. Accessed 13 August 2017. Oostende, Belgium: UNESCO/IOC Project Office for IODE. http://www.jcomm.info/WW12

Babanin, A. V., M. Onorato, and F. Qiao. 2012. Surface waves and wave-coupled effects in lower atmosphere and upper ocean. J. Geophys. Res. Oceans, *117*, C00J01. doi: 10.1029/2012JC007932

Babanin, A. V., and W. E. Rogers. 2014. Generation and limiters of rogue waves. Int. J. Ocean Clim. Syst., *5*, 39–49. doi: 10.1260/1759-3131.5.2.39

Belibassakis, K. A. 2007. A coupled-mode model for the scattering of water waves by shearing currents in variable bathymetry. J. Fluid Mech., *578*, 413–434. doi: 10.1017/S0022112007005125

Benjamin, T. B., and J. E. Feir. 1967. The disintegration of wave trains in deep water. Part 1. Theory. J. Fluid Mech., *27*, 417–430. doi: 10.1017/S002211206700045X

Bennis, A.-C., F. Ardhuin, and F. Dumas. 2011. On the coupling of wave and three dimensional circulation models: Choice of theoretical framework, practical implementation and adiabatic tests. Ocean Model., *40*, 260–272. doi: 10.1016/j.ocemod.2011.09.003

Belcher, S. E., and J. C. R. Hunt. 1998. Turbulent flow over hills and waves. Annu. Rev. Fluid Mech., *30*, 507–538. doi: 10.1146/annurev.fluid.30.1.507

Bingham H. B., and H. Zhang. 2007. On the accuracy of finite-difference solutions for nonlinear water waves. J. Eng. Math. *58*, 211–228. doi: 10.1007/s10665-006-9108-4

Chalikov, D., A. V. Babanin, and E. Sanina. 2014. Numerical modeling of 3D fully nonlinear potential periodic waves. Ocean Dyn., *64*, 1469–1486. doi: 10.1007/s10236-014-0755-0

Chalikov, D., and D. Sheinin. 1998. Direct modeling of one-dimensional nonlinear potential waves, *in* Nonlinear Ocean Waves, vol. 2, Advances in Fluid Mechanics Series 17, W. A. Perrie, ed. New York: Springer, pp. 207–258.

Chalikov, D., and D. Sheinin. 2005. Modeling extreme waves based on equations of potential flow with a free surface. J. Comput. Phys., *210*, 247–273. doi: 10.1016/j.jcp.2005.04.008

Chawla, A., and J. T. Kirby. 2002. Monochromatic and random wave breaking at blocking points. J. Geophys. Res. Oceans, *107*, 4-1–4-19. doi: 10.1029/2011JC001042

Clamond, D., and J. Grue. 2001. A fast method for fully nonlinear water-wave dynamics. J. Fluid Mech., *447*, 337–355. doi: 10.1017/S0022112001006000

Dommermuth, D. G., and D. K. P. Yue. 1987. A high-order spectral method for the study of nonlinear gravity waves. J. Fluid Mech., *184*, 267–288. doi: 10.1017/S002211208700288X

Donelan, M. A., A. V. Babanin, I. R. Young, and M. L. Banner. 2006. Wave-follower field measurements of the wind-input spectral function. Part II. Parameterization of the wind input. J. Phys. Oceanogr., *36*, 1672–1688. doi: 10.1175/JPO2933.1

Donelan, M. A., A. V. Babanin, I. R. Young, M. L. Banner, and C. McCormick. 2005. Wave-follower field measurements of the wind-input spectral function. Part I. Measurements and calibrations. J. Atmos. Ocean. Technol., *22*, 799–813. doi: 10.1175/JTECH1725.1

Donelan, M. A., M. Curcic, S. S. Chen, and A. K. Magnusson. 2012. Modeling waves and wind stress. J. Geophys. Res. Oceans, *117*, C00J23. doi: 10.1029/2011JC007787

Dysthe, K. B. 1979. Note on a modification to the nonlinear Schrödinger equation for application to deep water waves. Proc. R. Soc. Lond A Math. Phys. Sci., *369*, 105–114. doi: 10.1098/rspa.1979.0154

Gain, L. 1918. La prediction des houles au Maroc. Ann. Hydrogr., 65–75.

Garrett, C. 1976. Generation of Langmuir circulations by surface waves: A feedback mechanism. J. Mar. Res., *34*, 117–130.

Gelci, R., J. Cazale, and J. Vassal. 1956. Utilisation des diagrammes de propagation a la provision energitique de la houle. Bull. Inform. Comite Gen. Oceanogr. Etudes Cotes, *8*, 169–187.

Gelci, R., J. Cazale, and J. Vassal. 1957. Prevision de la houle. La method densites spectroangulaires. Bull. Inform. Camile Central Oceanogr. D'Etude Cotes, *9*, 416–435.

Ghantous, M., and A. V. Babanin. 2014. Ocean mixing by wave orbital motion. Acta Phys. Slovaca, *64*, 1–56. doi: 10.2478/apsrt-2014-0001

Gramstad, O., and A. V. Babanin. 2016. The generalized kinetic equation as a model for the nonlinear transfer in third-generation wave models. Ocean Dyn., *66*, 509–526. doi: 10.1007/s10236-016-0940-4

Gramstad, O., and M. Stiassnie. 2013. Phase-averaged equation for water waves. J. Fluid Mech., *718*, 280–303. doi: 10.1017/jfm.2012.609

Gravois, U. M., W. E. Rogers, and T. G. Jensen. 2012. A coupled model system for southeast Florida: Wave model validation using radar and in situ observations. Naval Research Laboratory (NRL) Report NRL/MR/7320-12-9355. Arlington, VA: Office of Naval Research, 45 p.

Groeneweg, J., and G. Klopman. 1998. Changes in the mean velocity profiles in the combined wave-current motion described in GLM formulation. J. Fluid Mech., *370*, 271–296. doi: 10.1017/S0022112098002018

Hanley, K. E., S. E. Belcher, and P. P. Sullivan. 2010. A global climatology of wind-wave interaction. J. Phys. Oceanogr., *40*, 1263–1282. doi: 10.1175/2010JPO4377.1

Hasselmann, K. 1962. On the non-linear energy transfer in a gravity-wave spectrum. Part 1. General theory. J. Fluid Mech., *12*, 481–500. doi: 10.1017/S0022112062000373

Hasselmann, S., K. Hasselmann, J. H. Allender, and T. P. Barnett. 1985. Computations and parameterizations of the nonlinear energy transfer in a gravity-wave spectrum. Part II: Parameterizations of the nonlinear energy transfer for application in wave models. J. Phys. Oceanogr., *15*, 1378–1391. doi: 10.1175/1520-0485(1985)015<1378:CAPOTN>2.0CO.2

Haus, B. K. 2007. Surface current effects on the fetch-limited growth of wave energy. J. Geophys. Res. Oceans, C112. doi: 10.1029/2006JC003924

Herbers, T. H. C., S. Elgar, and R. T. Guza. 1995. Generation and propagation of infragravity waves. J. Geophys. Res. Oceans, *100*, 24863–24872. doi: 10.1029/95JC02680

Hodur, R. M. 1997. The Naval Research Laboratory's Coupled Ocean/Atmosphere Mesoscale Prediction System (COAMPS). Mon. Weather Rev., *125*, 1414–1430. doi: 10.1175/1520-0493(1997)125<1414:TNRLSC>2.0.CO;2

Holthuijsen, L. H. 2007. Waves in Oceanic and Coastal Waters. Cambridge: Cambridge University Press. 387 p.

Iafrati, A., A. V. Babanin, and M. Onorato. 2013. Modulational instability, wave breaking and formation of large scale dipoles. Phys. Rev. Lett., *110*, 184504. doi: 10.1103/PhysRevLett.110.184505

Janssen, P. A. E. M. 2003. Nonlinear four-wave interaction and freak waves. J. Phys. Oceanogr., *33*, 863–884. doi: 10.1175/1520-0485(2003)33<863:NFIAFW>2.0CO;2

Janssen, P. A. E. M. 2004. The Interaction of Ocean Waves and Wind. Cambridge: Cambridge University Press. 300 p.

Jeffreys, H. 1924. On the formation of waves by wind. Proc. R. Soc. A, *107*, 189–206. doi: 10.1098/rspa.1925.0015

Jeffreys, H. 1924. On the formation of waves by wind. II. Proc. R. Soc. A, *110A*, 341–347.

Jenkins, A. D. 1989. The use of a wave prediction model for driving a near-surface current model. Dtsch. Hydrogr. Z., *42*, 133–149. doi: 10.1007/BF02226291

Jessup, A. T., C. J. Zappa, and H. Yen. 1997. Defining and quantifying microscale wave breaking with infrared imagery. J. Geophys. Res., *C*102, 23145–23153. doi: 10.1029/97JC01449

Kharif, C., E. N. Pelinovskii, and A. Slunyaev. 2009. Rogue Waves in the Ocean. New York: Springer, 16 p.

Kinsman, B. 1965. Wind Waves: Their Generation and Propagation on the Ocean Surface. Englewood Cliffs, NJ: Prentice-Hall, 76 p.

Komen, G. I., L. Cavaleri, M. Donelan, K. Hasselmann, S. Hasselmann, and P. A. E. M. Janssen. 1994. Dynamics and Modelling of Ocean Waves. Cambridge: Cambridge University Press, 554 p.

Kudryavtsev, V. N., V. K. Makin, and J. F. Meirink. 2001. Simplified model of air flow above the waves. Boundary Layer Meteorol., *98*, 155–171. doi: 10.1023/A:1018914113697

Lavrenov, I. V. 1998. The wave energy concentration at the Agulhas current of South Africa. Nat. Hazards, *17*, 117–127. doi: 10.1023/A:1007978326982

Lavrenov, I. V. 2003. Wind-Waves in Oceans: Dynamics and Numerical Simulations. New York: Springer,;77 p.

Longuet-Higgins, M. S. 1970a. Longshore currents generated by obliquely incident sea waves: 1. J. Geophys. Res., *75*, 6778–6789. doi: 10.1029/JC075i033p06778

Longuet-Higgins, M. S. 1970b. Longshore currents generated by obliquely incident sea waves: 2. J. Geophys. Res., *75*, 6790–6801. doi: 10.1029/JC075i033p06790

Longuet-Higgins, M. S., and R. W. Stewart. 1964. Radiation stresses in water waves: A physical discussion, with applications. Deep Sea Res. Oceanogr. Abstr., *11*, 529–562. doi: 10.1016/0011-7471(64)90001-4

Mellor, G. 2003. The three-dimensional current and surface wave equations. J. Phys. Oceanogr., *33*, 1978–1989. doi: 10.1175/1520-0485(2003)033<1978:TTCASW>2.0.CO;2

Mei, C. C. 1983. The Applied Dynamics of Ocean Surface Waves. New York: Wiley, 40 p.

Miles, J. W. 1957. On the generation of surface waves by shear flows. Part 1. J. Fluid. Mech., *3*, 185–204. doi: 10.1017/S0022112057000567

Miles, J. W. 1959. On the generation of surface waves by shear flows. Part 2. J. Fluid. Mech., *6*, 568–582. doi: 10.1017/S0022112059000830

Montagne, R. 1922. Le service de prediction de la houle au Maroc. Ann Hydrogr, 157–186.

Montiel, F., V. A. Squire, and L. G. Bennets. 2016. Attenuation and directional spreading of ocean wave spectra in the marginal ice zone. J. Fluid Mech., *790*, 492–522. doi: 10.1017/jfm.2016.21

Onorato, M., A. R. Osborne, M. Serio, and S. Bertone. 2001. Freak wave in random oceanic sea states. Phys. Rev. Lett., *86*, 5831–5834. doi: 10.1103/PhysRevLett.86.5831

Phillip, O. M. 1957. On the generation of waves by turbulent wind. J. Fluid Mech., *2*, 417–445. doi: 10.1017/S0022112057000233

Phillips, O. M. 1966. The Dynamics of the Upper Ocean. Cambridge: Cambridge University Press, 61 p.

Pinho, U. F., and A. V. Babanin. 2015. Emergence of short-crestedness in originally unidirectional nonlinear waves. Geophys. Res. Lett., *42*, 4110–4115. doi: 10.1002/2015GL063973

Plant, W. J. 1982. A relationship between wind stress and wave slope. J. Geophys. Res., *C87*, 1961–1967. doi: 10.1029/JC087iC03p01961

Rogers, W. E., A. V. Babanin, and D. W. Wang. 2012. Observation-consistent input and whitecapping-dissipation in a model for wind-generated surface waves: Description and simple calculations. J. Atmos. Oceanic Tech., *29*, 1329–1346. doi: 10.1175/JTECH-D-11-00092.1

Rogers, W. E., J. D. Dykes, and P. A. Wittmann. 2014. US Navy global and regional wave modeling. Oceanography, *27*, 56–67. doi: 10.5670/oceanog.2014.68

Shabani, B., A. V. Babanin, and T. E. Baldock. 2016. Observations of the directional distribution of the wind energy input function over swell waves. J. Geophys. Res. Oceans, *121*, 1174–1193. doi: 10.1002/2015JC011225

Shrira, V. I., and A. V. Slunyaev. 2014. Nonlinear dynamics of trapped waves on jet currents and rogue waves. Phys. Rev. E, *89*, 041002. doi: 10.1103/PhysRevE.89.041002

Smith, J. A. 2006. Wave-current interactions in finite depths. J. Phys. Oceanogr., *36*, 1403–1419. doi: 10.1175/JPO2911.1

The WISE Group, L. Cavaleri, J.-H. G. M. Alves, F. Ardhuin, A. Babanin, M. Banner, K. Belibassakisf, et al. 2007. Wave modeling: the state of the art. Prog. Oceanogr., *75*, 603–674. doi: 10.1016/j.pocean.2007.05.005

Tsagareli, K. N., A.V. Babanin, D. J. Walker, and I. R. Young. 2010. Numerical investigation of spectral evolution of wind waves. Part 1. Wind input source function. J. Phys. Oceanogr., *40*, 656–666. doi: 10.1175/2009JPO4345.1

Tulin, M. P., and T. Waseda. 1999. Laboratory observations of wave group evolution, including breaking effects. J. Fluid Mech., *378*, 197–232. doi: 10.1017/S0022112098003255

van der Westhuysen, A. J. 2012. Spectral modeling of wave dissipation on negative current gradients. Coast. Eng., *68*, 17–30. doi: 10.1016/j.coastaleng.2012.05.001

van der Westhuysen, A. J. 2016. Wave-current interaction, *in* Encyclopedia of Marine & Offshore Engineering, John Wiley & Sons, Ltd., doi: 10.1002/9781118476406.emoe085, 18p.

West, B., K. Brueckner, R. Janda, M. Milder, and R. Milton. 1987. A new numerical method for surface hydrodynamics. J. Geophys. Res., *92*, 11803–11824. doi: 10.1029/JC092iC11p11803.

Young, I. R. 1999. Wind Generated Ocean Waves. New York: Elsevier, 88 p.

Young, I. R., and Y. Eldeberky. 1998. Observations of triad coupling of finite depth wind-waves. Coast. Eng., *33*, 137–154. doi: 10.1016/S0378-3839(98)00006-4

Young, I. R., S. Zieger, and A. V. Babanin. 2011. Global trends in wind speed and wave height. Science, *332*, 451–455. doi: 10.1126/science.1197219

Yuen, H. C., and B. M. Lake. 1982. Nonlinear dynamics of deep-water gravity waves. Adv. Appl. Mech., *22*, 67–229. doi: 10.1016/S0065-2156(08)70066-8

Zakharov, V. E. 1966. The instability of waves in nonlinear dispersive media (in Russian). Zh. Eksp. Teor. Fiz. Pis'ma Red., *51*, 1107–1114.

Zakharov, V. E. 1967. The instability of waves in nonlinear dispersive media [Engl. Transl.]. Sov. Phys. JETP, *24*, 740–744.

Zakharov, V. E. 1968. Stability of periodic waves of finite amplitude on the surface of deep fluid [Engl. Transl.]. J. Appl. Mech. Tech. Phys., *2*, 190–194. doi: 10.1007/BF00913182

Zhang, F. W., W. M. Drennan, B. K. Haus, and H. C. Graber 2009. On wind-wave-current interactions during the Shoaling Waves Experiment. J. Geophys. Res. Oceans, *1142*, C01018. doi: 10.1029/2008JC004998.

Received: *7 March 2016*; revised: *10 October 2016*.

Editor's note: Contributions to *The Sea: The Science of Ocean Prediction* are being published separately in special issues of *Journal of Marine Research* and will be made available in a forthcoming supplement as Volume 17 of the series.