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起伏地表下地震波传播数值模拟方法研究进展

Progress in numerical simulation of seismic wave propagation under an undulating surface

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摘 要 起伏地表是地震数据的采集、处理和解释中普遍遇到的难题.起伏地表下的地震波传播数值模拟,对起伏地表观测的地震资料处理解释有重要意义.地震波场模拟和地震波走时场分别描述地震波的动力学和运动学信息,为研究地震波传播理论的两种重要途径.本文从地震波场和地震波走时场两方面回顾和总结了起伏地表下的地震波传播数值模拟方法的研究进展,并展示了该领域的一些最新研究成果,为使读者能从中找到突破点,为起伏地表这一勘探领域的经典难题做出贡献.

关键词 起伏地表;地震波传播;数值模拟;走时场

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Abstract Rough topography is very common and we have to deal with it during the acquisition, processing, and interpretation of seismic data. Modeling of seismic-wave propagation in medium with irregular topography is beneficial to process and interpret seismic data acquired by active and passive source seismology conducted in areas of interest such as mountain ranges and basins. Seismic wavefield and seismic wave traveltime field are two important ways to describe seismic wave propagation. In this paper, we try to give a basic review on the research progress of seismic wave propagation simulation method through the above two aspects, and show some new research results in this field, in the meaning of supplying a initial material so as to enable readers get some basic knowledge of the development obtained in the past and current, and find some breakthroughs then may make some contributions to the development or applications of new methods and technologies themselves.

Keywords undulating surface; seismic wave propagation; numerical simulation; traveltime

0 引 言

地球的表面常常是起伏或是不规则的,且实际的地球物理观测绝大部分是在这些地表崎岖的地方进行的(Robertsson and Holliger, 1997).例如,深地震测深中为了理解造山带、盆地等的形成机制,测线常常穿过这些地表剧烈起伏的地区(Teng *et al*., 1987, 2003; Zeng *et al*., 1995; Li and Mooney, 1998; Gao *et al*., 2000, 2005; Zhang and Klemperer, 2010).在油气(阎世信等, 2000; Yilmaz and Doherty, 2001; 张永刚, 2007; Lan and Zhang, 2011a, 2012; Ma and Zhang, 2014a, 2014b)与矿产资源(Farquharson *et al*., 2008; Lelièvre *et al*., 2009, 2012)地球物理勘探中,地球物理学家也遇到同样的问题.剧烈的地形起伏给地震探测工作提出了严峻的挑战.

起伏地表下的地震波传播数值模拟,不仅是对起伏地表观测的地震资料进行偏移和反演的基础,而且对后续处理结果的解释具有重要意义.地震波场和地震波走时场作为描述地震波传播的两种重要途径,常常通过对它们的模拟来研究

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地震波的传播规律和传播特征.下面我们就从地震波场和地震波走时场两方面来阐述起伏地表下的地震波传播数值模拟方法的研究进展.

1　起伏地表下的地震波场数值模拟方法研究进展

近年来,随着计算机技术的快速发展和计算能力的大幅提高,提出了一些有针对性的地震波场数值模拟方法.包括有限元法,伪谱法,边界元法,谱元法和有限差分法等.这些方法在模拟起伏地表的地震波场时各有优劣,下面对这些方法逐一介绍,表1简单的概括了这些数值方法在进行复杂地表的波场模拟时的优点和缺点.

表1　起伏地表下的地震波场数值模拟方法

Table 1　Methods for simulating seismic wavefields in media with an irregular surface

数值模拟方法	优点	缺点
有限元法	处理不规则地表方便	耗内存大,计算速度慢,成本高
谱元法	处理不规则地表方便	成本高
伪谱法	精度高,占用内存小	处理地表结构复杂或地表剧烈起伏以及地下结构复杂的情况时存在较大的误差
边界元法	处理不规则地表方便	不适用于地表速度变化较大情况
有限差分法	计算速度快,占用内存小,在模拟复杂介质中地震波传播时应用最为广泛	处理复杂地形比较困难

1.1　有限元法(FE)

有限元法基于变分原理,采用分段近似,基于三角形网格来剖分模型,该剖分保证了复杂地层形态模拟的逼真性(Rial *et al.*, 1992; Toshinawa and Ohmachi, 1992).但有限元法算法复杂,计算速度慢,占用内存和运算量均较大.为此,发展了一些有限元和其他方法结合的混合方法.例如,Moczo等用离散波数方法模拟震源激发和下部介质中地震波的传播,而通过有限元方法来模拟沿起伏地表传播的波(Moczo *et al.*, 1997; Galis *et al.*, 2008);张美根等(2002)研究了各向异性弹性波有限元正演系统的精度和效率问题,提出了一种透射加衰减的组合人工边界方案(吸收边界条件);杨顶辉等(2002)基于双相各向异性介质模型,推导了双相各向异性介质中弹性波传播的动力学方程及其Galerkin变分方程和有限元运动方程,对双相PTL介质和双相各向同性介质中的弹性波传播进行了数值模拟;黄自萍等(2004)提出了一种有限元和有限差分(FE-FD)藕合的区域分裂方法,该方法克服了单纯用差分方法对区域的依赖性,而同样的精度所需的计算量比有限元法小,利用这种方法成功地模拟了起伏地表地震波的传播.但是这类方法在两种方法的交界处易产生人为反射,吸收边界也不易处理.刘有山等(2013)采用稀疏存储的显式有限元三角形网格波场模拟方法,减少了计算中的内存;并采用PML边界条件,提高了吸收边界效果.

1.2　伪谱法(PS)

伪谱法又叫虚谱法,是一种逼近空间微分的方法,利用傅里叶变换将波场函数表示为傅里叶级数的展开形式,在时间一波数域或频率域中求解波动方程,精度高,占用内存小(Tessmer *et al.*, 1992; Tessmer and Kosloff, 1994; Nielsen *et al.*, 1994).进入上世纪90年代之后,该方法有了飞跃式的发展.石玉梅(1995)给出了流体饱和多孔隙介质中弹性波传播数值模拟的伪谱法.张文生等(1998)用伪谱法进行了二维横向各向同性介质波动方程的正演模拟,特别是对边界吸收问题作了有效的处理.Furunura等(1995)首次针对伪谱法提出了反周期扩展边界方法;Takenaka和王彦宾等(Takenaka *et al.*, 1999, Wang *et al.*, 2000)利用伪谱法分别计算了球对称全球模型和具有垂向速度梯度的沉积盆地模型中地震波的传播问题.之后不久,Wang and Takenaka (2001)利用不连续网格傅立叶伪谱多域方法模拟了区域地球模型中弹性波的传播,接下来Wang等(2001)利用伪谱法模拟了二维柱坐标下全地球模型中地震波的传播.赵志新等(2003)给出了非均匀介质中地震波传播数值模拟的错格实数傅立叶伪谱法.赵景霞等(2003)用伪谱法来模拟曲线网格的二维声波方程,提高了计算效率;Tessmer and Kosloff (1994)利用坐标变换的思想,采用傅立叶变换法与有限差分法相结合的方法对起伏地表的弹性波波场进行了模拟.由于傅里叶变换是基于整个时间域或空间域的,改变空间中的某一点的值,就会改变频率域中的所有值,因此每一点的微分结果都要受到计算域中其它点的影响并且存在众所周知的Gibbs效应(Landsberg, 1978).实际上,求导运算是一种局部运算.对于空间物性剧烈变化的情形,这种处理显然是不合适的,即在处理地表结构复杂或地表剧烈起伏以及地下结构复杂的情况时存在较大的误差(Tessmer *et al.*, 1992).

1.3　边界元法(BE)

边界元法也叫边界积分法,该方法将散射波场用地表的一个半解析的积分来表示,其中积分项中Green函数一般在频率波数域中计算(Banerjee and Butterfield, 1981).边界元法在研究起伏地表地震波传播时使用较多(Campillo and Bouchon, 1985; Bouchon *et al.*, 1989; Sánchez-Sesma and Campillo, 1991, 1993; 符力耘和牟永光, 1994; Durand *et al.*, 1999; Sánchez-Sesma *et al.*, 2006).与其它的模拟方法相比,BEM的优势主要在于:

(1)不涉及体积离散,只需要在求解域的边界上进行离

散化处理,降低了问题的维数;

(2)能够精确地描述地下不规则界面的几何特征;

(3)对于无界域问题,自动满足远场辐射条件.

传统的BEM虽有上述优点,但边界元方法的半解析性质决定了该方法不适用于地表速度变化较大情况,而实际情况是,由于后期地质作用造成浅部地层速度变化剧烈,这限制了边界元法方法的实际应用(Bouchon *et al*., 1995).传统BEM的另一个不足就是它的计算效率问题,这是由它在计算过程中会产生不对称矩阵所致.符力耕等(1994)采用单元长度随介质速度和计算频率变化的变单元算法以及自动剖分单元等技术而提高了计算效率. Zhou等(2008, 2009a, 2009b)提出了一个局部化的边界积分-波数离散方法.由于在这种方法中所采用的逆矩阵大小只与地表起伏部分的采样数成比例,因此在保证精度的前提下,计算效率有了很大提高.

1.4　谱元法(SE)

近年来提出的谱元法是将有限元法和谱展开法相结合的方法.它的基本思想是选取以正交多项式表示的形函数,在各个单元上通过配置点插值,提高解的收敛速度(Komatitsch and Vilotte, 1998; Komatitsch and Tromp, 1999, 2002).在该方法中,计算范围被分成许多子区域,每个子区域中的解可以表示成截断的契比雪夫或勒让德多项式级数的乘积,用伽辽金方法求解正交问题的变分格式.谱元法是弹性波方程在空间近似的高阶有限元方法,具有有限元法处理不规则地表比较方便的特征. Dimitri. Komatisch等(Komatitsch and Vilotte, 1998; Komatitsch and Tromp, 1999, 2002; Komatitsch *et al*., 2000)用谱元法模拟了二维空间起伏地表下的弹性波场,讨论了复杂地表引起的各种干扰,并利用该方法模拟了按正弦规律起伏的海底反射、透射等.国内学者近年来也开始陆续用谱元法开展了地震波场的数值模拟研究.林伟军等(2005)在基于Lagrange多项式展开的SEM中引入了逐元技术,降低了内存需求和计算量.后来林伟军等(2005)又详细论述了Chebyshev谱元法的基本理论和相应的数学公式.王童奎等(2007)也进行了横向各向同性介质中的地震波场数值模拟. Liu等(2014)把有限元法和谱元法做了详细对比,发现当使用相同数量的插值节点时,有限元法较谱元法精度高,然而有限元法较难推广到高阶,另外,四边形网格上的有限元法和谱元法较三角形网格上的精度高且计算速度快.该研究从某种程度上改变了人们对有限元和谱元方法的认识,使人们可以在数值模拟时根据实际需要客观地选择方法.

1.5　有限差分法(FD)

有限差分法是通过有限差分算子将波动方程离散化,以差分代替微分,将微分方程化为代数方程组,然后求解该线性代数方程组以获得微分方程的数值解(Mitchell and Griffiths, 1980).差分算子是一个空间局部的算子,在空间域具有较高的分辨率,可以很好地适应剧烈变化的地下介质情况(Wong, 1982; Jih *et al*., 1988; Frankel and Vidale, 1992; Hestholm and Ruud, 1994, 1998; Robertsson, 1996; 张剑锋, 1998; Hestholm *et al*., 1999; Oprsal and Zahradnik, 1999; Hayashi *et al*., 2001; Hestholm, 2003; Gao and Zhang, 2006; Zhang and Chen, 2006; 董良国等, 2007; 刘永霞等, 2007; 张华等, 2007; 张金海等, 2007; Lombard *et al*., 2008).在频率域中,有限差分算子的分辨率很低,仅适合于相对较简单的地质模型.但算法的计算速度快,占用内存小,在模拟复杂介质中地震波传播时应用最为广泛,但该方法的一个主要缺陷是处理复杂地形比较困难.为此,Tessmers等(1992)提出了一种新的思路,即通过坐标变换将具有起伏地表的模型及弹性波方程变换到新的具有水平地表的坐标系中,在新坐标系中求解弹性波方程,时间上用差分法,空间上横向用傅立叶法,纵向采用契比雪夫方法计算波场对空间的导数,使用Gottlieb等(1982)提出的吸收边界和自由边界. Hestholm等(Hestholm and Ruud, 1994, 1998; Hestholm *et al*., 1999)借鉴了Tessmer等的思想,通过坐标变换将起伏地表转换成水平地表后,完全用差分法求解弹性波方程,在适应地表起伏的同时提高了计算效率. Tessmer, Hestholm等处理起伏地表的曲化平思想的实质是对垂直方向上的网格单元进行压缩或拉伸. Tessmer等人考虑的是一种平均高差分配(相似)准则,而Hestholm等人采用的是一种绝对高差分配(相似)准则.但是,他们的方法都需要建立地表的函数模型,且模型的好坏对计算结果有很大影响.裴正林(2004)运用交错网格任意偶数阶精度差分格式求解一阶速度-应力方程;然后采用将零速度法和广义虚像法相结合的方法来处理自由边界,并在自由边界上采用四阶精度差分格式,成功地对起伏地表下的弹性波波场进行了模拟.兰海强等(2011)通过引入坐标变换和流体力学中的贴体网格,将笛卡尔坐标系中的波动方程变换到曲线坐标系,从而把物理空间中的不规则模拟区域转化为计算空间的规则区域,然后采用一种稳定的、显式的二阶精度的有限差分方法离散(曲线坐标系中的)弹性波方程,从而发展了一种模拟起伏地表下地震波场的数值方法.该方法可以灵活适应复杂边界,且实现简单,计算稳定.之后,他们又对横向各向同性介质(Lan and Zhang, 2011b)、流体填充裂缝介质(Lan and Zhang, 2012)的地震波场进行了模拟研究.图1展示了一个二维半圆形凹陷模型,图2展示了兰海强等模拟的该模型中的VTI介质波场的地震记录(兰海强等, 2011).从图中可以看出,由于凹陷的存在,qP波和瑞雷面波的能量在凹陷右侧明显减小;在直达qP波之后观察到由瑞雷面波散射转换成的次生qP波(RqPf);同理,在瑞雷面波之前也观察到由于qP散射转换的次生瑞雷面波(qPRf);还有一些反射的

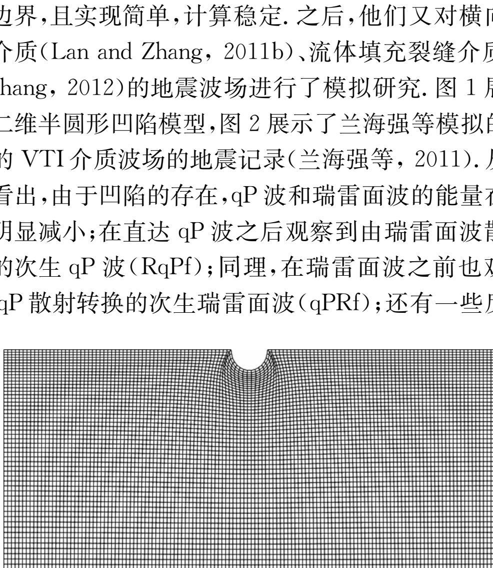

图1　半圆形凹陷模型示意图

Fig.1　A semi-circular shape depression topography model

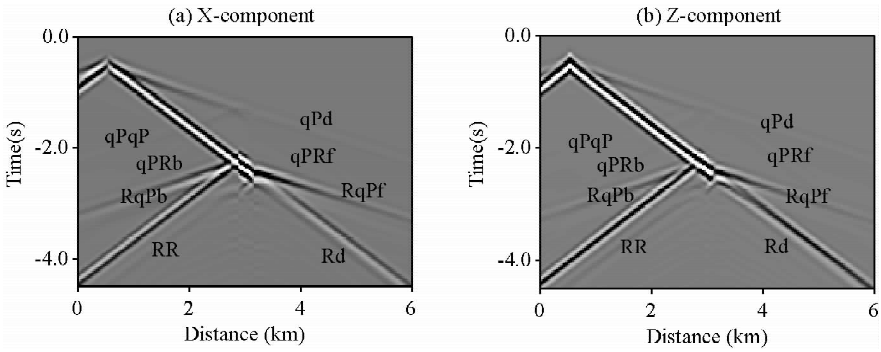

图 2　半圆形凹陷模型中波场位移 x 分量(a)和 z 分量(b)记录.图中 qPd，Rd 分别表示 qP 波衍射波和瑞雷面波衍射波；qPRf，qPRb 分别表示 qP 波发生散射,转换为瑞雷面波向前和向后传播;RqPf，RqPb 分别表示瑞雷面波发生散射,转换为 qP 波向前和向后传播;qPqP 表示 qP 波反射波;RR 表示瑞雷面波反射波.

Fig. 2　Seismograms for the semi-circular shape depression topography model. Symbols mean the following: (qPd) qP wave diffracts to qP wave; (Rd) Rayleigh wave diffracts to Rayleigh wave; (qPRf) qP wave scatters to Rayleigh wave and propagates forward; (qPRb) qP wave scatters to Rayleigh wave and propagates backward; (qPqP) qP wave reflectes to qP wave; (RqPf) Rayleigh wave scatters to qP wave and propagates forward; (RqPb) Rayleigh wave scatters to qP wave and propagates backward; (RR) Rayleigh wave reflectes to Rayleigh wave.

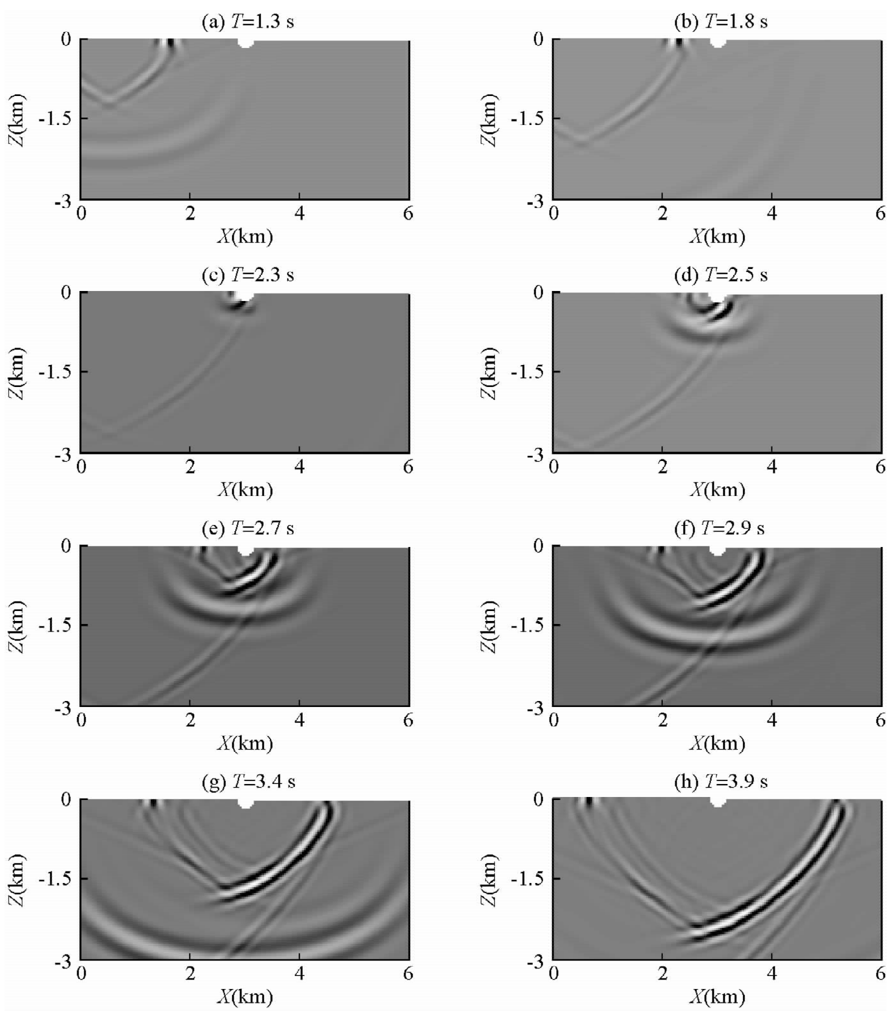

图 3　半圆形凹陷模型中波场位移 z 分量在不同时刻的快照

Fig. 3　Snapshots of the vertical component of the wavefield at different propagation times for the semi-circular shape depression topography model

瑞雷面波(qPRb, RR)和 qP 波(qPqP, RqPb). 凹陷的边缘处,由于地形突变引起体波和瑞雷面波发生强烈散射. 由于面波的波长较小,因此瑞雷面波产生的散射要明显强于体波,这表明这样的陡凹陷模型能明显地阻碍瑞雷面波的传播. 图 3 为垂直分量的波场快照,可观察到 qP 波,qSV 波,瑞雷面波以及连接 qSV 波和沿地表传播的 qP 波的首波;在 2.3 s时(图 3c),瑞雷面波刚刚到达小山丘表面,开始产生反射波和转换波. 同样可观察到由小凹陷产生的反射 qSV 波.

2 起伏地表下的地震波走时场数值模拟方法

2.1 基于程函方程数值解的地震波走时场

地震波走时是地震波的一个重要属性参数,在地震勘探正反演研究和生产实践中有着重要的应用(Yilmaz and Doherty, 2001). 事实上,很多种用于对地球内部结构和物质组成进行成像和探测的方法都要用到地震波走时,如用于构造成像的 Kirchhoff 积分偏移(Gray and May, 1994; Symes *et al.*, 1994)和用于速度场反演的地震层析成像(Hole, 1992; Zelt and Smith, 1992; Zhang *et al.*, 2000; Aki and Richards, 2002; Zhang *et al.*, 2010).

在地震勘探数据处理与成像中,Kirchhoff 积分偏移是一种直接映射记录波场到模型空间的工业成像方法,该方法的核心之一就是计算复杂介质中的地震波走时(Berkhout, 1984). 一方面,对于每一个成像点都要计算它的绕射时间曲面,计算量很大;另一方面,偏移速度场通常比较复杂,这对二维地震波走时的算法提出了较高的要求,走时算法的速度和精度直接决定着成像方法的应用范围和效果(Yilmaz and Doherty, 2001).

同样,在地震勘探和地震深部探测中,地震层析成像已经取得了广泛的应用成果,它既可以用来解决局部地下构造问题(Zhang *et al.*, 2000, 2005, 2009, 2010; Zhang *et al.*, 2013),也可以用来解决区域构造问题(Widiyantoro and Van Der Hilst, 1997; Gorbatov *et al.*, 2000, 2001),还可以用来解决全球性问题(Hole, 1992; Zelt *et al.*, 2001; 徐涛等, 2007; Xu *et al.*, 2014). 层析成像是通过若干个源与检波器之间的走时来反演速度场的方法,该方法预先给定一个初始速度模型,由此速度模型可以计算出一组理论走时,然后通过对比理论走时和实际走时,不断调整速度模型,最终得到满意的反演结果.

因此,地震波走时计算技术在地球物理的正演和反演理论研究以及实际生产中都有着很重要的应用. 因此,对地震波走时计算技术的研究意义重大.

传统的计算走时的方法为射线方法,其原理是沿程函方程的特征方向,即射线方向求解走时,经插值计算后得到地下规则网格点上的地震波走时(Cerveny *et al.*, 1977). 但是,对于复杂构造的地质模型,传统的射线追踪方法可能会产生阴影区 (Cerveny *et al.*, 1977; Cerveny, 2001; Xu *et al.*, 2006, 2010, 2014; 徐涛等, 2004; 李飞等, 2013),为了满足处理实际地震资料的需要,研究合适的地震波走时算法并进一步提高地震波走时计算的速度和精度十分重要. Vidale 基于扩张波前的思想开创性地提出一种用有限差分方法来近似程函方程(Vidale, 1988, 1990),但该方法采用扩张矩形来追踪波前,在一定的速度分布情况下,计算出的旅行时并不是最小,且在处理强速度界面时会出现不稳定现象. Qin 等改进了 Vidale 的方法,尽可能沿扩张的波前面来计算旅行时,方法和 Vidale 基本相同但他考虑到了因果关系,首先寻求上一个近似波前面的旅行时全局极小点,然后向外扩张(Qin *et al.*, 1992);但该方法计算机实现较困难,大部分时间要用于寻找全局极值,效率不高. Van Trier 等人将迎风有限差分法引入解程函方程,大大的提高了差分格式的稳定性(Van Trier and Symes, 1991). Sethian 等提出了一种称之为快速推进的方法(Fast marching method, FMM)(Sethian, 1996, 1999, 2001; Sethian and Popovici, 1999),该方法利用迎风差分格式求解局部程函方程,采用窄带延拓重建旅行时波前,利用堆选排技术保存旅行时,将最小旅行时放在堆的顶部. 该方法显著缩短了寻找极小值的时间,计算量由波前扩展法的$O(N^a)$减少到 O(N · logN)(logN 由堆的排序算法产生),其中 N 是节点数. 近年来,很多学者对快速推进法进行了推广和应用(Fomel, 1997; Alkhalifah and Fomel, 2001; Kim, 2002; Rawlinson and Sambridge, 2004a, 2004b, 2005; 孙章庆等, 2009). 最近,Zhao 提出了一种称之为快速扫描的方法(Fast sweeping method, FSM)(Zhao, 2005)用于求解一阶双曲型偏微分方程,并指出该方法的计算量为 O(N),且证明了该算法的单调性和稳定性. 实际上快速扫描法和快速推进法都是求解相同的离散方程,且都是基于因果关系沿程函方程的特征方向求解,但前者更强健且对高阶方程更灵活,效率也一般较后者高(Zhang *et al.*, 2006; 刘一峰和兰海强, 2012; 兰海强等, 2012a, 2012b).

2.2 起伏地表下的地震波走时场

前面提到的计算地震波走时的方法都是基于水平地表的. 研究复杂地表条件下地震波走时的计算问题,对于在这些地区进行的反射地震偏移成像(Gray and May, 1994; Symes *et al.*, 1994)、走时层析反演均有非常重要的意义(Morelli and Dziewonski, 1987; Hole, 1992; Piromallo and Morelli, 1997; Badal *et al.*, 2004; Trampert *et al.*, 2004; Zhao, 2004; Trampert and van der Hilst, 2005; Wang, 2011). 到目前为至,可以处理起伏地表的地震波走时计算方法都是基于非规则网格的(Fomel, 1997; Kimmel and Sethian, 1998; Sethian, 1999; Sethian and Vladimirsky, 2000; Rawlinson and Sambridge, 2004a; Qian *et al.*, 2007a, 2007b; Kao *et al.*, 2008; Lelièvre *et al.*, 2011). 然而,一般的,无论是在网格剖分阶段还是走时计算阶段,基于不规则网格的方法都比基于规则网格的方法需要更多的计算量,而且目前绝大多数的基于不规则网格的方法只能处理剖分的三角形夹角为锐角的情况(Lelièvre *et al.*, 2011). 另外,目前常用的很多地震波场数值模拟方法都是基于规则网格的,这使在这些网格上的走时计算变得愈加重要,因为基于以上的正演结果进行偏移时需要这些走时信息(Fornberg, 1988; Tessmer *et al.*, 1992; Hestholm and Rudd, 1998; Zhang and Chen, 2006; Appelo and Petersson, 2009). 因此,基于规则网格剖分的起伏地表下的走时计算具

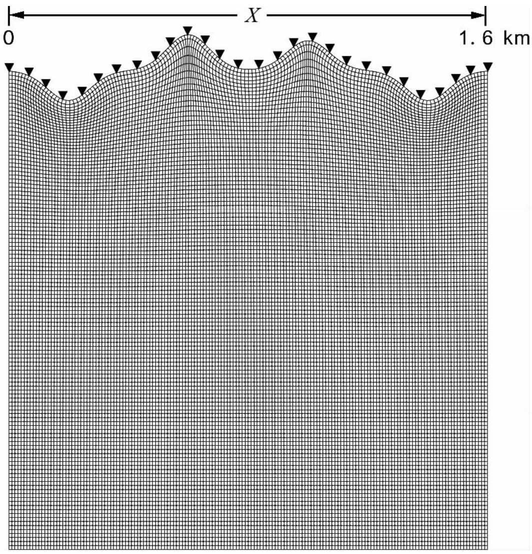

图 4 剖分剧烈起伏地表模型的贴体网格

Fig. 4 The boundary conforming grids for discretizing a model with a rather irregular surface

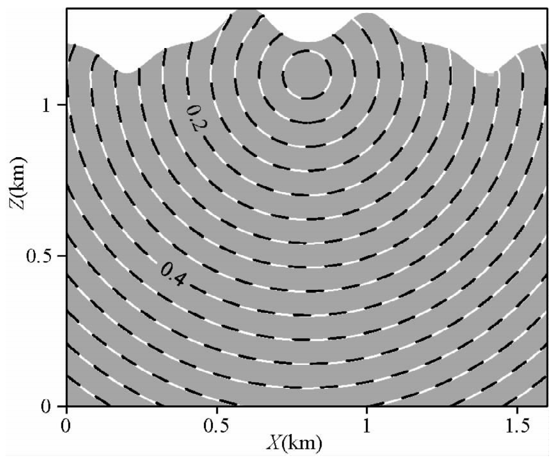

图 5 剧烈起伏地表模型中计算的初至走时等值线

震源坐标为(0.8 km, 1.1 km);白色实线表示解析解,黑色虚线表示数值解,走时等值线的单位为秒.

Fig. 5 Traveltime contours (in seconds) for the severely irregular surface model with the source is at (0.8 km, 1.1 km). The black dashed and white solid lines denote the numerical and analytical solutions, respectively.

有重要意义.最近,Sun 等(2011)提出了一种在规则网格上求解起伏地表下地震波走时的方法,该方法的关键是在起伏地表附近采用非均一的网格间距且通过引入地表点、地表上点、界面点、界面下点等概念使其适于用 FMM 求解.然而该方法需要对不规则地表的走时进行插值处理,这样势必会引入误差.兰海强等创新地提出了与地形相关的地震波的程函方程并给出了相应的求解方法,提供了一种计算起伏地表初至走时的全新的思路和方法(Lan and Zhang, 2013a).他们通过坐标变换推导出了曲线坐标系的程函方程,解决了经典的程函方程求解起伏地表走时的困难,将不规则区域上的经典的程函方程变换为规则区域上的各向异性的程函方程.此时,不能用求解各向同性程函方程的较流行的快速推进法直接求解该方程.为此,兰海强等引入求解 Hamailton-Jacobi 方程的 Lax-Friedrichs 快速扫描法来其求解地形相关的程函方程,数值实例表明该方法稳定、准确,但效率较低.为了得到满足精度要求的结果,常常需要对求解区域进行精细剖分,这严重影响了求解效率,难以满足生产实践的需求.为此,兰海强等引入数学上用于求解守恒率方程的高阶加权本质非振荡格式(WEND)来替换 Lax-Friedrichs 快速扫描法中的低阶差分格式,发展了求解与地形有关的程函方程的高效算法(Lan and Zhang, 2013b).该方法显著地提高了计算的效率,在较粗的网格上需要较少的迭代次数就能得到满足精度要求的解.最近兰海强等将该方法推广到了各向异性异性介质(Lan *et al*., 2014).图 4 为一个由两个山丘和两个凹陷的组合而成的地表剧烈起伏的模型.模型大小为 1.6 km×1.32 km,地表在 1.1~1.32 km 之间起伏.图 5 展示了兰海强等用他们发展的方法计算的该模型中的初至波走时场(Lan and Zhang, 2013a).

3 结 论

起伏地表是地震数据的采集、处理和解释中普遍遇到的难题.起伏地表下的地震波传播数值模拟,对起伏地表观测的地震资料处理解释具有重要意义.本文从地震波场和地震波走时场两方面回顾和总结了起伏地表下的地震波传播数值模拟方法的研究进展,并展示了该领域的一些最新研究成果,为使初入该领域的学者能对起伏地表下地震波场和走时场的模拟的发展历程和现状有一定了解,进而能取其所长,避其所短,为实际资料的处理解释和新方法的研究创新做出贡献.

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